Proceedings of the Sixth International Soil Correlation Meeting (VI ISCOM): Characterization, Classification, and Utilization of Cold Aridisols and Vertisols

Montana, Idaho, and Wyoming, United States, and Saskatchewan, Canada

August 6-18, 1989

March, 1991

Edited by J.M. Kimble
Soil Conservation Service
Soil Management Support Services
USDA

Soil Conservation Service
USDA

Soil Management Support Services
USAID

Correct Citation:
Kimble, J.M. 1990. Proceedings of the Sixth International Soil Correlation Meeting (VI ISCOM) - Characterization, Classification, and Utilization of Cold Aridisols and Vertisols. USDA, Soil Conservation Service, National Soil Survey Center, Lincoln, NE.

Copies of this publication may be obtained from:

National Leader World Soil Resources
Soil Management Support Services
Soil Survey Division
USDA-SCS
P.O. Box 2890
Washington D.C. 20013 USA

National Leader for Soil Classification
Soil Survey Division
USDA-SCS
P.O. Box 2890
Washington D.C. 20013 USA

Proceedings of the Sixth International Soil Correlation Meeting (VI IS

Characterization, Classification, and Utilization of Cold Aridisols and

Montana, Idaho, and Wyoming, United States, and Saskatchewan, Cana

August 6-18, 1989
March, 1991

Organized by:

USDA, SCS, SMSS, Lincoln, NE and Washington, DC
Agriculture Canada
USDA, SCS, Montana, Idaho, and Wyoming
USDI, BLM, Montana and Wyoming
University of Saskatchewan, Saskatoon, Saskatchewan, Canada

Sponsored by:

USDA, SCS, Washinton, D.C.
USAID, Washington, D.C.

Editor
J.M. Kimble

Workshop Committees

Steering Committee

H. Eswaran, National Leader SMSS, SCS, Washington, DC
J. Witty, National Leader, Soil Classification, SCS, Washington, DC
Juan Comerma, Chairman of ICOMERT
Ahmet Osman, Chairman of ICOMID

Organizing Committee

J. Witty, National Leader, Soil Classifiction, SCS, Washington, DC (Chairman)
J. Kimble Research Soil Scientist, SMSS, SSIV, NSSC, SCS, Lincoln, NE
T.D. Cook, Soil Mangement Specialist, SMSS, SCS, Washington, DC
A. Mermut, University of Saskatchewan, Saskatoon, Saskatchewan, Canada
D. Acton, Agriculture Canada, Saskatoon, Saskatchewan, Canada

Organized by

USDA, SCS, SMSS, Lincoln, NE and Washington, DC
Agriculture Canada
USDA, SCS, Montana, Idaho, and Wyoming
USDI, BLM, Montana and Wyoming
University of Saskatchewan, Saskatoon, Saskatchewan, Canada

Preface

The Sixth International Soil Correlation Meeting (VI ISCOM) was organized by the Soil Mangement Support Services in conjunction with the Soil Survey Division, Soil Conservation Service, the SCS Soil Staffs and BLM Soils Staffs in Montana, Idaho, and Wyoming, Agriculture Canada, Saskatoon, Saskatchewan, Canada, and the University of Saskatchewan, Saskatoon, Saskatchewan, Canada. The meeting was held August 6-18, 1989.

The meeting addressed cold Aridisols and Vertisols. Proposed Keys developed by the chairmen of the respective committees were tested in the field and discussed. Modifications of the proposals were presented and new proposals developed. Based on the discussion, a revision of the Vertisol order was developed and submitted to John Witty. Proposals to change the Aridisol order are still under consideration.

One of the more interesting and far-reaching proposals was the one to drop the Aridisol order and include soils which now fall in it in the other orders, based on their gentic horizons. This proposal created a very active discussion in which very strong feelings were expressed. We thank all of the tour participants for such frank discussions. Without them, the meeting could not have been as successful as it was. Having a large number of participants with a common interest interacting as a group helps develop new ideas and a better understanding of old ones.

This *Proceedings* was assembled to allow others to have the benefit of the material presented on the tour by the various authors. It represents a current reference to the latest thinking on Aridisols and Vertisols. The idea to drop the Aridisol order was not accepted at this time, but it did generate a great deal of discussion and made many of the participants look at the idea. In fact, many thought it had merit. It even lead to a proposal to also drop the Mollisol order, which may have been suggested in jest but has merit if we consider how the other orders are defined. As editor, I feel both of these topics will arise again and cause many soil scientists to examine the concepts of both Aridisols and Mollisols, two orders which seem not to follow the ideas of *Soil Taxonomy*.

J.M. Kimble, Editor

Acknowledgements

The editor wishes to thank all of the authors who prepared manuscripts and presented them to the tour participants and then sent them in to be published in this Proceedings. Thanks also are due to all of the participants who took part in the peer review of the manuscripts. Without the efforts of all concerned, the maunscripts would not have ended up in the form which they did. Special thanks is also given to Maria Lemon, Ph.D., of the Editor Inc., for completing an English edit of all of the manuscripts and for preparing the proceedings in its final format.

Thanks also go to Kristen Stuart, Anthony Flores, and Nancy Martinez for handling all of the mailings for the manuscripts' technical edits and for correcting the final manuscripts. Without their time and efforts, this Proceedings would not have been completed.

Special thanks also are given to G. Durnbush, Director, Midwest National Technical Center, for making so many of the National Soil Survey Centers employees available to take part and help in conducting the tour and in preparing the manuscripts for this Proceedings. The State conservationists and the soils staffs from the SCS offices in Montana, Idaho, and Wyoming are thanked for all of their efforts, as are the many other soil scientists who helped make the VI ISCOM a sucess.

Similar thanks go to the State BLM diredctors in Montana and Wyoming and their soil staffs for active and enthusiastic support for the tour. We also thank Dr. Acton, Agriculture Canada, and Dr. Mermut, University of Scakatchewan, for their assistance in all phases of ISCOM VI.

We also wish to recognize the steering committe members for their encouragement and guidance. Special thanks go to Dr. Juan Comerma for preparing the Vertisol proposals and to Dr. A. Osman for preparing the Aridisol proposals. We also would like to regonize both Dr. J. Witty and Dr. H. Eswaran for their contributions to both proposals. Without the efforts of Dr. Witty and his staff, the many proposals put forth would not be considered, and much careful consideration and testing is required before the necessary changes can be made to Soil Taxonomy to reflect the contributions of such a wide and diverse group of participants and committe members.

We are genuinely indebted and grateful to the all these individuals and staffs; however, we do not want to overlook the many others, too numerous to name, who contributed to making the ISCOM such a sucess. We gratefully recognize their contributions.

The Editor,
for all of the organizing committee

Contents

Classification of Vertic Intergrades: Macromorphological and Micromorphological Aspects

W.A. Blokhuis*, L.P. Wilding, and M.J. Kooistra[1]

Abstract

Differentiation between typic and vertic subgroups of Alfisols, Inceptisols, and Mollisols has been based mainly on the shrink-swell potential of the soil material, expressed as a coefficient of linear extensibility (COLE) and/or potential linear extensibility (PLE). Clay plasticity index and liquid limit also have been used as differentiae. Other soil characteristics are often different between vertic and typic subgroups, or between vertic subgroups and Vertisols, but these are insufficiently significant to be used as differentiating criteria.

Micromorphological data do not consistently support any separation between vertic intergrades and Vertisols or between vertic and typic subgroups in other soil orders. Some forms of plasma separations and voids may seem to be characteristic of Vertisols, but they are not unique and can not be used as differentiae.

This paper proposes to differentiate between two kinds of intergrades: firstly, soils that fulfill the present requirements on COLE/PLE, have vertic characteristics in the surface soil but lack such characteristics in the B horizon because of insufficient wet-dry cycles to generate shear failure, and, secondly, soils with a vertic soil structure at some depth but lacking vertic characteristics in the surface soil.

Introduction: Present and Former *Soil Taxonomy* Criteria

Vertisols occur in the KEY to Soil Orders of *Soil Taxonomy* (Soil Survey Staff, 1987) subsequent to Histosols, Spodosols, and Oxisols, as "Other soils that:

1. Do not have a lithic or paralithic contact, petrocalcic horizon, or duripan within 50 cm of the surface;
   and
2. After the soil to a depth of 18 cm has been mixed, as by plowing, have 30 percent or more clay in all subhorizons to a depth of 50 cm or more;
   and
3. Have, at some time in most years unless irrigated or cultivated, open cracks at a depth of 50 cm that are at least 1 cm wide and extend upward to the surface or to the base of the plow layer or surface crust;
   and
4. Have one or more of the following:
   a. Gilgai;
   b. At some depth between 25 cm and 1 m, slickensides close enough to intersect;
   c. At some depth between 25 cm and 1 m, wedge-shaped natural structural aggregates that have their long axes tilted 10 to 60 degrees from the horizontal."

If the requirement on gilgai is waived (in accordance with ICOMERT proposals), criterion 4 is essentially one on 'vertic structure'. We use the term 'vertic structure' when wedge-shaped or parallelepiped structural aggregates and/or intersecting slickensides are present. If a pedon satisfies both the requirements on cracking (3) and on soil structure (4), those on soil depth (1) and on clay content (2) normally will be satisfied as well.

<u>Vertic subgroups</u> have been defined in the Orders Alfisols, Aridisols, Entisols, Inceptisols, Mollisols and Ultisols (Soil Survey Staff, 1987). Vertic subgroups differ from typic subgroups in the same great group by having:

1. Cracks at some period in most years that are 1 cm or more wide at a depth of 50 cm, that are at least 30 cm long in some part, and that extend upward to the soil surface or to the base of an Ap horizon (in some vertic Alfisols: to the base of an albic horizon);
   and
2. More than 35 percent clay in horizons that have total thickness of 50 cm;
   and

[1]Agricultural University, Wageningen, The Netherlands; Texas A & M University, USA; The Winand Staring Centre for Integrated Land, Soil and Water Research, Wageningen, The Netherlands.

*Corresponding author

3. A coefficient of linear extensibility (COLE) of more than 0.05, 0.07 or 0.09 (depending on the soil moisture regime of the great group) in a horizon or horizons at least 50 cm thick, and a potential linear extensibility (PLE) of 6 cm or more in the upper 150, 125 or 100 cm, respectively, of the soil, or in the whole soil if a lithic or paralithic contact is deeper than 50 cm but shallower than 150, 125 or 100 cm, respectively (in some great groups there is no requirement for COLE).

The mimimum COLE requirements are 0.09 for udic and aquic, 0.07 for ustic, and 0.05 for xeric and aridic soil moisture regimes. Different COLE requirements have been introduced in order to reflect different moisture gradients: if the soil moisture changes are large or frequent, a smaller COLE would produce the same movement in the soil as a higher COLE would do in an environment where these changes are small or infrequent (Soil Survey Staff, 1975).

The criteria on cracking are practically the same for vertic subgroups and for Vertisols, but there are differences in the requirements on clay content in certain sections of the profile. Swell-shrink characteristics must be shown in Vertisols in a specific subsoil structure, whereas in vertic subgroups requirements on COLE and PLE indicate a potential to swell-shrink.

The *Soil Taxonomy* concept of vertic intergrades is one of soils that lack sufficient evidence of soil movement to meet the definition of Vertisols but that have high shrink-swell potential. In most vertic subgroups, wet-dry cycles are insufficiently expressed to generate shear failure. A vertic subgroup may develop into a Vertisol with time, and under changing environmental conditions. One could think of the following examples:

1. A floodplain, in which the surface soils have sufficiently developed to produce structural aggregates and cracks, but where the subsurface soil is still in an 'Entisol-' or 'Inceptisol-stage,' perhaps not yet fully ripened (assuming adequate COLE).
2. A smectitic clay soil under aridic conditions, where the sparse and erratic rainfall has created sufficient dry/wet alternations in the surface soil to generate cracks but has seldom moistened the subsurface soil to such an extent that swelling pressures and shear failure are generated.

If vertic soils are potential Vertisols, it is not surprising that there are many common properties. Bartelli and McCormack (1976) emphasize the great similarity between Vertisols and soils in vertic subgroups in such characteristics as instability because of swelling, clay mineralogy, COLE value, clay content, plasticity index, liquid limit, 15-bar (1500kPa) water and cation exchange capacity. Both Vertisols and vertic soils belong to group CH (clayey soils with high liquid limits) in the Unified Soil Classification System (DeMent and Bartelli, 1969).

In *Soil Taxonomy* (Soil Survey Staff, 1975) Lithic Vertic and Paralithic Vertic subgroups were recognised in Mollisols and Inceptisols. These had a lithic or paralithic contact between 25 and 50 cm depth and horizons totalling 25 cm or more in thickness that had either more than 35% montmorillonitic clay or COLE >0.05 or 0.07 (depending on soil moisture regime). In the 3rd edition of the KEYS (Soil Survey Staff, 1987) these subgroups have been waived.

Vertic Integrades: The Need for a Wider Concept

No provision is made in *Soil Taxonony* for the many Alfisols, Mollisols, and Inceptisols that have a vertic structure in some part of the solum - usually accompanied by relatively high COLE values and high clay contents - but that do not meet the requirements on cracking and therefore can not be placed in a vertic subgroup.

In clayey soils it is the combination of cracking and a vertic structure that characterizes Vertisols. There would be some logic in defining soils that have either the morphological surface soil characteristics (i.e., cracking) or the morphological subsurface soil characteristics (i.e., a vertic structure) as vertic intergrades. In *Soil Taxonomy*, however, only the first category is recognized - with the additional requirement (COLE, PLE) that the soil material has a potential to develop a vertic structure. A wider concept of vertic intergrades that includes non-cracking clay soils with a vertic structure in the subsoil is worth considering.

When scanning soil reports, excursion guides, and the like on vertic 'intergrades', one finds that the term 'vertic' often has been applied rather loosely to soils with cracks, soils with pressure faces, soils with slickensides. However, there is always one vertic symptom: either the cracking or the vertic subsoil structure. And

this seems, from the point of view of macromorphology, a reasonable entrance.

Of the nine pedons described as vertic subgroups in India (Kooistra, 1982), none is convincingly vertic according to the present *Soil Taxonomy* criteria. The name vertic has been applied when there are cracks and/or when there are features of a vertic structure. COLE values given for some of these pedons meet the requirements for vertic subgroups.

Vertic subgroups of Alfisols, Inceptisols, and Mollisols in Texas, USA, were described by Hallmark et al. (1986). Profile and site descriptions and analytical data, including COLE, were given. Cracks have been recorded in only three of the eleven vertic intergrades, but one has to realise that cracks were described only when they were apparent at the time of field description. COLE was always above the critical value. Slickensides, pressure faces, and parallelepipeds - alone or in combination - occurred at some depth in ten of the eleven pedons.

If rules are to follow practice, then these observations underline the need for a wider concept of the vertic subgroup than the present one.

Macro- and Micromorphological Considerations

There are many suggestions in the literature that the obvious differences in macromorphology between Vertisols and vertic subgroups (according to the present definitions) are matched with differences in micromorphology. However, if one tries to generalize the observations that have been made, there is the problem that vertic subgroups may, or may not, have a vertic structure.

The nine benchmark soils of India, described as vertic subgroups (Kooistra, 1982), had weakly to moderately developed vosepic and/or skelsepic plasmic fabric. Only two pedons in this group had a distinct vertic macrostructure and masepic plasmic fabric in some part.

Plasmic fabric of Vertisols described in the same study was usually moderately to strongly developed vo-skelsepic, whereas four Vertisols had, in addition, masepic plasmic fabric. Skelsepic plasmic fabric only occurred in connection with coarse fragments over 4 mm diameter.

Dasog et al. (1987) described four pedons with vertic properties in Saskatchewan, Canada: two Vertisols and two Argic Vertic Cryoborolls. The main differences between Vertisols and vertic intergrades were in macromorphology (slickensides, versus prismatic structure with illuviation argillans) and in micromorphology (masepic and lattisepic plasmic fabric were common throughout most of the solum in Vertisols, whereas they were few in vertic subgroups, and confined to the BC horizon). There were minor differences in other characteristics (CEC, clay content, clay mineralogy, COLE, pH, liquid limit, plasticity index), and, if these were taken into consideration as well, next to the macro- and micromorphological features, Vertisols differed clearly from Vertic Borolls. However, it was difficult to single out one or two of these properties as diagnostic criteria.

Bui and Mermut (1988) measured the total length, the area percentage (total length per unit area. in cm/cm2), and the orientation (in 10 degree-intervals) of planar voids, using an image analyser, in three Vertisols from India, four from Ghana, and one from Saskatchewan, as well as three vertic subgroups from Saskatchewan. The vertic soils had columnar or prismatic macrostructure and most planar voids were oriented vertically, whereas in Vertisols the dominant orientation of planar voids was subhorizontal and oblique.

Mermut et al. (1988) state that vertic subgroups lack planar vosepic, skelsepic, and masepic plasmic fabric.

Nettleton et al. (1983) tried to find micromorphological parameters for turbation. They studied eight Vertisols and six soils in vertic subgroups (two Torrifluvents and four Hapludalfs) in thirteen counties across the United States. Physical and chemical characteristics (including linear extensibility) were similar in Vertisols, Torrifluvents, and the Bt horizons of the Hapludalfs (other horizons of the Hapludalfs were not investigated). There were differences in macromorphology: vertic subgroups had cracks, but lacked intersecting slickensides; and in micromorphology: the Torrifluvents had a silasepic plasmic fabric, the Hapludalfs had an omnisepic plasmic fabric, and the Vertisols had skelsepic plasmic fabric in the upper and masepic in the lower part of the solum. One of the Hapludalfs had a few slickensides and areas with masepic plasmic fabric. Note that vosepic plasmic fabric was not described in any of the soils discussed.

Masepic plasmic fabric, with minor areas with skelsepic and vosepic plasmic fabric, was characteristic for Chromic Pelloxererts described by Nettleton and Sleeman (1985).

Masepic plasmic fabric was common, next to vosepic and skelsepic plasmic fabric, in the Branyon series (microlow), a Udic Pellustert in Texas (Hallmark et al.,1986).

In calcareous lower Bw horizons of Vertisols, the plasmic fabric was crystic, but decalcified thin sections revealed mostly masepic plasmic fabrics (Wilding and Drees, 1989).

The above citations suggest that certain micromorphological features, notably masepic plasmic fabric - but to a smaller extent also the vosepic, skelsepic, and lattisepic types - and planar voids are characteristic of Vertisols and are rare, weakly developed or absent in vertic subgroups. However, other observations show that these features also are found in vertic subgroups and in non-vertic soils, and in some cases are more strongly developed in non-vertic soils than in related Vertisols.

Many of the soils reported by McCormack and Wilding (1974; 1975), Ritchie et al. (1974), Smith and Wilding (1972), Smeck et al. (1981), Stahnke et al. (1983), Rabenhorst and Wilding (1986), Sobecki and Wilding (1983), and Wilding and Tessier (1988) had skelsepic, masepic, and lattisepic plasmic fabrics. Some had, in addition, vosepic plasmic fabric. The soils studied were Vertisols, vertic subgroups in other orders, and non-vertic clayey, fine-loamy and loamy soils. Most of the non-vertic soils belonged to aquic suborders or subgroups. They included Aeric and Mollic Ochraqualfs, Aeric Fragiaqualfs, Typic Argiaquolls, and Aquic and Typic Paleustalfs.

Yerima et al. (1987) found skelsepic and masepic, next to insepic, mosepic, and argillasepic plasmic fabric, in a Vertic Argiustoll in El Salvador.

Nettleton and Sleeman (1985) described some non-vertic clayey soils that had plasma modifications like Vertisols, viz. masepic, lattisepic, and vosepic plasmic fabric.

Blokhuis et al. (1989) gave examples of ripening clay sediments that had a stronger development of vosepic and masepic plasmic fabric than was usually found in Vertisols and in vertic subgroups. Unistrial and omnisepic plasmic fabrics were also common in such sediments.

Linear Extensibility

Studies by Nettleton et al. (1969) showed that soils with argillic horizons that had omnisepic, lattisepic, and masepic plasmic fabrics had COLE values in the range 0.05-0.06, whereas in similar soils that had mosepic plasmic fabric COLE values were mostly around 0.03. Nettleton et al. (1983) found that Vertisols with masepic plasmic fabric usually had COLE 0.06-0.16, but some Vertisols without masepic plasmic fabric had COLE 0.04-0.06, i.e., below the requirements on COLE for vertic subgroups.

McCormack and Wilding (1975) and Yule and Ritchie (1980) showed that expansive soils continue to swell at moisture tensions below 1/3-bar tension and continue to shrink at moisture tensions above 15 bar. COLE, therefore, is an index for swelling and shrinking between limits, not a measure of maximum swell and shrink. Also, COLE is only a potential index of shrink-swell. McCormack and Wilding (1975) found that a confinement stress equivalent to 10% of the swelling pressure would reduce the percentage total swell by 50%. Hence, actual shrink-swell displacement is strongly conditional on overburden (soil thickness).

Discussion

Diagnostic Criteria for Vertisols, Vertic Subgroups, and Non-vertic Soils

The micromorphology of vertic subgroups strongly overlaps with Vertisols and non-vertic soils. Some authors working in specific areas could differentiate between Vertisols and vertic intergrades, but even so the differences found were in grade of development and relative abundance rather than in type of plasmic fabric or voids. We found suggestions of correlations (see above), the most distinct one being that between masepic plasmic fabric and a vertic macrostructure. This observation however, is, not of great help, as vertic subgroups and non-vertic soils may or may not have a vertic macrostructure and, hence, may or may not have masepic (and vosepic, lattisepic) plasmic fabric. Skelsepic plasmic fabric is a less reliable parameter for stress, as it is influenced by amount and size of sand particles and coarse fragments.

At first sight the picture gets even more confused if we turn to non-vertic soils. Nettleton and Sleeman (1985) found that some non-vertic clayey soils had plasma modifications like Vertisols, viz., masepic, lattisepic, and vosepic plasmic fabric. Blokhuis et al. (1989) gave examples of ripening clay sediments that had a stronger development of vosepic and masepic plasmic fabric than is usually found in Vertisols and in vertic subgroups. Unistrial and omnisepic plasmic fabrics are also common in such sediments. These authors also found that soil strength fail-

ure - that obviously occurs mainly along major slickensides - is not correlative with microshear in the soil matrix: vosepic plasmic fabric appeared to be most strongly developed along the narrowest of planar voids.

Nettleton et al. (1983) gave the following interpretation of micromorphological features, based on their own work and that of others:
- Planar voids are an expression of depositional and/or stress-strain history of the soils;
- Silasepic plasmic fabric indicates absence of strain; omnisepic, skelsepic and lattisepic plasmic fabrics indicate a more or less balanced three-dimensional strain. These forms would occur in vertic subgroups;
- Masepic plasmic fabric is an expression of strain; it is typical of Vertisols, especially in lower horizons.

McCormack and Wilding (1974) interpreted masepic and lattisepic plasmic fabrics as stress failures. Wilding and Tessier (1988) reported on the microscopic and submicroscopic basis of shear failure.

The occurrence of vosepic plasmic fabric in many Mollisols and Alfisols that show no macormorphological evidence of shear (see e.g. Stahnke et al., 1983; Ritchie et al., 1974; Sobecki and Wilding, 1983, and Smith and Wilding, 1972) is probably the result of plastic deformation rather than shear failure.

Brewer (1964) suggested that vosepic plasmic fabric is the result of forces just insufficient to cause shearing of the soil mass, whereas masepic plasmic fabric is produced by shearing. Jim (1986) found, in experimental studies, that planar voids and associated vosepic plasmic fabric resulted from shearing, whereas masepic plasmic fabric could be a precursor of vosepic plasmic fabric, prior to shearing and formation of a planar void. Field observations by Blokhuis et al. (1989) were in support of Jim's hypothesis.

Another interpretation of strain-related forms of plasmic fabric is that they result from tensile stress, not from shear stress. Tensile stress could explain the strong development of vosepic and masepic plasmic fabrics in ripening muds (Augustinus and Slager, 1971; Koenigs, 1984). Shrinkage of a saturated clay soil on drying can be regarded as a compression of the soil fabric by an all-round pressure that is quantitatively equivalent to the internal soil water suction (Towner, 1961).

Finally, one should realize that some methods of thin section preparation create soil stress and could enhance the development of shear-related forms of plasmic fabric or accentuate plasmic fabrics that were present in the natural soil. This may be the case when relatively wet clay soils (with pF 3 or less) are air-dried, or when epoxy resins are used for impregnation, requiring heating to 60-80 degrees Celsius.

The conclusion of the above discussion must be that the micromorphology does not provide us with suitable parameters to distinguish between Vertisols, vertic subgroups, and non-vertic soils. The question now arises: what other parameters are available?

There is, firstly, the macromorphology, with two major aspects: vertic structure and cracking. Vertic structure readily can be ascertained in the field. Quantification of aspects of soil structure could be considered but is difficult and perhaps not necessary. Vertic structure is not diagnostic for vertic subgroups according to the present *Soil Taxonomy* definition.

In the present definitions (Soil Survey Staff, 1987) the requirements on cracking are largely the same for Vertisols and for vertic subgroups. Reversible cracking is an obvious feature of Vertisols, although these cracks can be obscured by a surface mulch, or an Ap, or in a moist soil.

Linear extensibility expressed as COLE or as PLE was shown to indicate clearly whether or not a soil material had a potential for shrink-swell. Vertisols can be considered as soils that have realized this potential by developing a vertic structure. One would expect Vertisols to have at least the same COLE as vertic subgroups, and this was confirmed in many observations discussed above. The fact that COLE is a potential index for shrink-swell and not a measure for actual soil displacement - that is strongly conditional on overburden, as we have stated earlier - does not detract from its value as a differentiating characteristic between vertic and non-vertic soils.

Soils that formerly were recognised as Lithic or Paralithic Vertic subgroups (Soil Survey Staff, 1975) cannot now be named vertic, as they have a lithic or paralithic contact at less than 50 cm depth. One consequence of this is that many of the "shallow black soils" of the Deccan plateau in India, which have distinct vertic properties, cannot be classified either as a Vertisol or as a vertic subgroup, although they have distinct

vertic properties that are decisive for a specific land use. In fact, these are probably Vertisols that have been truncated by soil erosion after ages of cultivation.

If the soil material with high shrink-swell potential occurs deep in the profile, it would not normally be considered diagnostic in *Soil Taxonomy*. However, for engineering purposes this property would be a major constraint (Cook et al., 1988) that might be taken into account if the taxonomy is to serve areas outside agriculture.

Proposals to Amend the Definition of Vertic Subgroups

The present definition of vertic subgroups centers around cracking, COLE/PLE, and the presence of one or more horizons with a high clay percentage. The fact that no provision was made for the presence of a vertic structure at some depth in soils other than Vertisols has led several soil scientists to propose modifications.

Dasog et al. (1987) and Mermut et al. (1988) have suggested that, if the complete micromorphology and, if possible, other soil characteristics as well are taken into consideration, it will be possible to differentiate at least between Vertisol and vertic subgroup. We think, however, that this 'central concept' approach cannot provide us with parameters that can be applied world-wide.

Cook et al. (1988) proposed to confine the definition of vertic subgroups to the shrink-swell potential of the material and the requirement that the potential is present within a certain depth range. This proposal conforms, apart from details, to the present *Soil Taxonomy* requirements on COLE and PLE but waives the requirements on clay content and cracking.

Klich et al. (in press) proposed that vertic intergrades should have, within a defined depth zone (i.e. within 1.25m from the soil surface), a subhorizon with a minimum thickness (i.e. 30 cm) that has either a vertic structure or a minimum PLE (i.e. 5 cm). Cracks may or may not be present in soils that exhibit subsurface vertic conditions, and therefore these authors propose that the cracking requirement be waived. This proposal is similar to an earlier one by Graham and Southard (1983), who suggested that soils with significant shrink-swell activity (i.e. having a vertic structure) in subsoils that lie close to the surface probably should be allowed in vertic subgroups, regardless of surface cracking.

In the revised FAO/Unesco system of soil classification (FAO-Unesco-ISRIC, 1988) the term 'vertic properties' is used "in connexion with clayey soils which at some period in most years show one or more of the following: cracks, slickensides, wedge-shaped or parallelepiped structural aggregates, that are not in a combination, or are not sufficiently expressed, for the soils to qualify as Vertisols." Vertic soil units in Major Soil Groupings other than Vertisols have 'vertic properties'. This approach, although lacking in accuracy of definition, is attractive as it is based entirely on field morphology. It does not include poorly drained 'potential Vertisols' that do not (yet) show surface cracks.

In the French soil classification system (CPCS, 1967), vertic characteristics are recognised at a level comparable to subgroups in *Soil Taxonomy*; these may have any - but not all - of the characteristics of Vertisols.

Conclusions

Considering the above discussion, we propose to distinguish between two types of intergrades:

1. Those that fulfill the present requirements on COLE (e.g., over a thickness of at least 50 cm) and on PLE (e.g., in the upper 100 cm), that have cracks, but that lack a vertic structure. Desiccating clayey muds that develop irreversible cracking must be excluded from this concept.
2. Those that have a vertic structure at some defined depth (a requirement on COLE/PLE is not required as these values will be in the Vertisol range in the section with vertic structure).

Although the categories overlap, the former centre around potential Vertisols (cf. the present *Soil Taxonomy* definition of vertic subgroups), and the latter covers the soils that exhibit shear failure in the subsoil. The first category has the cracking features of a Vertisol, and the second normally has no cracks or cracks that are too small or too shallow.

Requirements on clay content perhaps could be waived: soils with either high COLE or a vertic structure in a section of the profile usually have clay contents over 30 or 35% in that depth range.

One could drop the COLE requirement or drop PLE and specify depth and thickness of layers that should have a specified mimimum COLE.

Literature Cited

Augustinus, P GEF, and S. Slager. 1971. Soil formation in swamp soils of the coastal fringe of Surinam. Geoderma 6:203-211.

Bartelli, L.J., and D.E. McCormack. 1976. Morphology and pedologic classification of swelling soils. Transportation Research Record 568. National Res. Council, Washington D.C.; 8 pp.

Blokhuis, W.A., M.J. Kooistra, and L.P. Wilding, 1989. Micromorphology of cracking clayey soils (Vertisols). Int. Working Meeting on Soil Micromorphology, San Antonio, Texas, July 10-15, 1988.

Brewer, R. 1964. Fabric and mineral analysis of soils. Wiley, New York etc., 470 pp.

Bui, E.N., and A.R. Mermut. 1988. Orientation of planar voids in Vertisols and soils with vertic properties. Soil Sci. Soc. Am. J. 52:171-178.

Cook, T.D., H. Eswaran, and J. Kimble. 1988. Vertic intergrades in Soil Taxonomy. Trans. Int. Workshop Swell-Shrink Soils, October 24-28, 1988, NBSS & LUP, Nagpur, India, pp. 45-47

CPCS, 1967. Classification des Sols. Travaux Commission de Pedologie et de Cartographie des Sols 1963-1967. 96 pp. Mimeo.

Dasog, G.S., D.F. Acton, and A.R. Mermut. 1987. Genesis and classification of clay soils with vertic properties in Saskatchewan. Soil Sci. Soc. Am. J. 51:1243-1250.

DeMent, J.A., and L.J. Bartelli. 1969. The role of vertic subgroups in the comprehensive soil classification system. Soil Sci. Soc. Am. Proc. 33:129-131.

FAO-Unesco-ISRIC. 1989. FAO-Unesco Soil Map of the World, Revised Legend. World Soil Resources Report 60, Food and Agriculture Organization of the United Nations, Rome, 119 pp.

Graham, R.C., and A.R. Southard. 1983. Genesis of a Vertisol and an associated Mollisol in Northern Utah. Soil Sci. Soc. Am. J. 47:552-559.

Hallmark, C.T., L.T. West, L.P. Wilding, and L.R. Drees. 1986. Characterization on data for selected Texas soils. Texas Agr. Expt. Station, Dept. of Soil and Crop Sciences, Soil Characterization Laboratory, College Station TX 77843.

Jim, C.Y. 1986. Experimental study of soil microfabrics induced by anisotropic stresses of confined swelling and shrinking. Geoderma 37:91-112.

Klich, I., L.P. Wilding, and A.A. Pfordresher (in press). Close-interval spatial variability of Udertic Paleustalfs in East-Central Texas.

Koenigs, F.F.R. 1981. Discussion on paper by L.P. Wilding, Development of structural and microfabric properties in shrinking and swelling clays. In J. Bouma and P.A.C. Raats (ed) Proceedings of the ISSS Symposium on Water and Solute Movement in Heavy Clay Soils. Int. Inst. Land Reclamation and Improvement, Publ. 37, Wageningen, The Netherlands, p. 22.

Kooistra, M.J. 1982. Micromorphological analysis and-characterization of 70 benchmark soils of India. Basic reference set. Neth. Soil Survey Inst., 778 pp.; in 4 volumes.

McCormack, D.E., and L.P. Wilding. 1974. Proposed origin of lattisepic frabic. In: G.K. Rutherford (Editor), Soil Microscopy. 4th Int. Working Meeting on Soil Micromorphology. The Limestone Press, Kingston, Ontario, pp. 761-771.

McCormack, D.E., and L.P. Wilding. 1975. Soil properties influencing swelling in Canfield and Geeburg soils. Soil Sci. Soc. Am. Proc. 39:496-502.

Mermut, A.R., J.L. Sehgal, and G. Stoops. 1988. Micromorphology of swell-shrink soils. Trans. Int.Workshop Swell-Shrink soils, October 24-28, 1988. NBSS & LUP, Nagpur, India, pp. 127-144.

Nettleton, W.D., K.W. Flach, and B.R. Brasher. 1969. Argillic horizons without clay skins. Soil Sci. Soc. Am. Proc. 33:121-125.

Nettleton, W.D., F.F. Peterson, and G. Borst. 1983. Micromorphological evidence of turbation in Vertisols and soils in vertic subgroups. In: P.Bullock and C.P.Murphy (Editors), Soil Micromorphology. Vol.2: Soil Genesis. AB Academic Publ., Berkhamsted, Hrts., United Kingdom, pp. 441-458.

Nettleton, W.D., and J.R. Sleeman. 1985. Micromorphology of Vertisols. In: L.A.Douglas and M.L.Thompson (Editors). Soil Micromorphology and Soil Classification. Soil Sci. Soc.Am. Special Publication no. 15, pp. 165-196.

Rabenhorst, M.C., and L.P. Wilding. 1986. Pedogenesis on theEdwards Plateau, Texas: II. Formation and occurrence of diagnostic subsurface horizons in a climosequence. Soil Sci. Soc. Am. J. 50:687-692.

Ritchie,A., L.P. Wilding, G.F. Hall, and C.R. Stahnke. 1974. Genetic implications of B horizons in Aqualfs of Northeastern Ohio. Soil Sci. Soc. Am. Proc. 38:351-358.

Smeck, N.E., A. Ritchie, L.P. Wilding. and L.R. Drees. 1981. Clay accumulation in sola of poorly drained soils of Western Ohio. Soil Sci. Soc. Am. J. 45:95-102.

Smith, H., and L.P. Wilding. 1972. Genesis of argillic horizons in Ochraqualfs derived from fine-textured till deposits of Northwestern Ohio and Southeastern Michigan. Soil Sci. Soc. Am. Proc. 36:808-815.

Sobecki, T.M., and L.P.Wilding. 1983. Calcic horizon distribution and soil classification in selected soils of the Texas coast prairie. Soil Sci. Soc. Am. J. 47:707-715.

Soil Survey Staff. 1975. Soil Taxonomy: a basic system of soil classification for making and interpreting soil surveys. USDA Agr. Handb. 346. U.S.Govt.Print.Office, Washington DC, 754 pp.

Soil Survey Staff. 1987. Keys to Soil Taxonomy (third printing). SMSS Technical Monograph No. 6, Ithaca, NY.

Stahnke, C.R., L.P. Wilding, J.D. Moore, and L.R. Drees. 1983. Genesis and properties of Paleustalfs of North Central Texas: morphological, physical and chemical properties. Soil Sci. Soc. Am. J. 47:728-733.

Towner, G.D. 1961. Influence of soil-water suction on some mechanical properties of soils. J. Soil Sci., 12:180-187.

Wilding, L.P., and L.R. Drees, in press. Removal of carbonates from thin-sections for microfabric interpretations. Int. Working Meeting on Soil Micromorphology, San Antonio, Texas, July 10-15,1988.

Wilding, L.P., and D. Tessier. 1988. Genesis of Vertisols: shrink-swell phenomena. In: Larry P.Wilding and Ruben Puentes (ed), Vertisols: their distribution, properties, classification and management. Techn.Monograph 18, Texas A & M University, Printing Center, pp. 55-81.

Yerima, B.P.K., L.P. Wilding, F.G. Calhoun, and C.T. Hallmark. 1987. Volcanic ash-influenced Vertisols and associated Mollisols of El Salvador: physical, chemical and morphological properties. Soil Sci. Soc.Am. J. 51:699-708.

Yule, D.F., and J.T. Ritchie. 1980. Soil shrinkage relationships of Texas Vertisols: I. Small cores. Soil Sci. Soc. Am. J. 44:1285-1291.

Desertification: Concept, Evaluation, Status

T.G.Boyadgiev[1]

Abstract

This paper considers desertification as a comprehensive expression of natural or induced processes which destroy the equilibrium of the soil, vegetation, air, and water. Desertification is a continuous process going through several stages before reaching the final one, which is an irreversible change. The process occurs in arid, semi-arid, and sub-humid areas.

The processes leading to desertification are considered to be degradation of the vegetative cover, water and wind erosion, salinization, reduction of soil organic matter, soil crusting and compaction, and accumulation of substances toxic to plants or animals. For each of these processes it is necessary to quantify the status, rate, inherent risk, and hazards of desertification.

To assess and map desertification, concrete socio-economic, bio-climatic, and physico-geographic data are necessary. If these data are not available, the desertification can be assessed by using mathematical models applying the interpretative approach. In this case the desertification hazards can be assessed as a function of the soil status, vulnarability of land to desertification processes (water action, wind action, salinization), and animal and population pressure on the land.

The data presented in the paper indicate that over 95 percent of the territory of Africa is covered with soils having adverse properties for one or more cultures and that the negative anthropogenic impact can lead to rapid land degradation.

The processes leading to desertification on the territory of the USSR are mainly vegetation degradation and salinization.

Introduction

The concept of desertification hazards is based on the following formulation (FAO/UNEP, 1983):

1. Desertification is a result of simultaneous action of natural mechanisms and mechanisms provoked by the influence of man and animals, but it can be restricted or stopped only through man's activities.
2. Desertification is a continuous process which passes through many stages before reaching the final one, which is an irreversible change.
3. Desertification occurs in arid, semi-arid, and sub-humid zones, and its intensity is greatest in marginal zones.

The basic processes of desertification are degradation of the vegetative cover, water erosion, wind erosion, salinization and alkanization, reduction of the organic matter, disturbance of soil physical properties (crusting and compaction), and accumulation of substances toxic for plants, animals, and men. It is assumed that the influence of climatic elements (drought and intensive rainfall, strong winds, high evapotranspiration, etc.) and anthropogenic activity leads much more quickly to desertification of areas with poor and shallow soils than of those having deep and fertile soils.

For an accurate evaluation of desertification hazards, as outlined in FAO/UNEP (1983) methodology, concrete socioeconomic, bio-climatic, and physicogeographic data are required in order to determine the status, rate, and inherent risk for each basic process. When such data are lacking or incomplete (as in developing countries) and the evaluation is urgently required, then mathematical models are used, applying the interpretative approach. In such cases, desertification is evaluated according to the following categories (FAO/UNEP/ESRI, 1984):

1. Anthropogenic influence: It is assumed that this influence has the strongest effect on destruction of the vegetative cover and reduction of soil fertility.
2. Land vulnerability to the water erosion, wind erosion, and salinization.
3. Soil status at the time of the degradation evaluation.

[1]N. Poushkarov Institute of Soil Science and Yield Programming.

Anthropogenic Effect

The continuous growth of population, as compared to the static character of the natural resources (soil, water, vegetation) and the restrictions of land biological productivity, is considered by many experts as proof of the growing influence of the demographic factor on the land and therefore as an indicator of soil degradation and desertification hazard. In the FAO/UNEP concept this obvious fact takes a quantitative expression and allows, even though not fully, a substantiation of the anthropogenic influence. To that aim, the following relations are used:

- Number of people/population supporting capacity - for assessment of demographic pressure.
- Number of animals/Livestock carrying capacity - for assessment of livestock pressure.
- The number of people and animals are taken from the annual statistics.

To establish the population supporting capacity, the following data are used: area by crop kinds, yield of each crop, potential annual production, and corresponding value of calories. The quantity of calories divided by the population's needs per year determines the amount of demographic pressure per unit area, and this value divided by the population is used as an indicator of the demographic factor effect on the environment. It is considered that this effect is zero to slight if the ratio people/population supporting capacity is less than 1,0, and very strong when the ratio is greater than 12.

To calculate the livestock carrying capacity for a given territory data on the total quantity of vegetation mass and the quantity of consumable fodder in kg per unit area are used. This amount, divided by the animals' requirements per year (2% of live weight per day), provides an idea of the degree of livestock pressure per unit area. The ratio of the number of animals/livestock carrying capacity is used as a quantitative expression of the animal pressure on the environment. This pressure is zero to slight when the ratio is less than 0,4, and very strong when it is greater than 4,2. These limit values are 2,5 times lower than those of the demographic factor, which shows that the method places greater weight on the influence of animals on land degradation.

In cases when no specific data on bio-mass are available, a formula can be used. Its calculation is based on the quantity of rainfall for the respective bio-climatic zone and on the soil factors restricting land productivity. Thus, for example, for the Mediterranean basin, Le Houerou (1977) proposes the following formula: consumable fodder (kg/ha) = 2,17P - 103,7, where P = mean annual precipitation (mm). This climatic potential is decreased by 25-50%, depending on soil properties unfavorable for development of plants (salinity, alkalinity, acidity, profile thickness, high gypsum and carbonates content, coarse texture, low fertility, etc.).

Land Vulnerability

Here is included the natural vulnerability of a given territory to water erosion, wind erosion, and salinization.

A large amount of related information exists, particularly about water erosion and salinization. This information, however, does not allow us to establish continuous trends for the speed of the process or the hazard of reaching irreversable degradation. It is assumed that the inclusion of available information in a system of mathematical models will allow us to outline the hazard zones of desertification.

The most important elements which the vulnerability assessment models include those discussed below.

For Water Erosion

For water erosion, the most important elements are:

- Aggressivity of precipitation with regard to natural vegetation and cultivated lands;
- Topographic effect;
- Coefficient related to the effects of texture, water permeability, and humus content of the soil;
- Coefficient related to rock weathering in various bio-climatic zones.

The values obtained from the combination of these elements can provide a quantitative expression of the water erosion hazard. It is considered as a zero to slight hazard if the value is lower than 1,9 and very great if this value is greater than 9,5.

Water erosion hazard is greatest for medium textured soils developed on a slope under tropical conditions. The ratio of texture to land slope is shown in Table 1. These data show that more than half of the land is medium textured and has rolling to mountainous slopes, and consequently is vulnerable to water erosion.

For Wind Erosion

The data about wind erosion hazard assessment are very limited and this made it necessary to rely on the findings of scientists that have studied the dependencies existing among the following elements:

- wind velocity and transportation of soil particles;
- mean wind velocity and percentage of active winds;
- movement of soil particles and moisture;
- soil erodibility and texture;
- status of soil surface and wind action.

The establishment of figure values for each one of these relationships permits a quantitative determination of the wind erosion hazard. This hazard is zero to slight when the index is lower than 3 and very great when it is greater than 15.

Judging by the light texture of the soils, which are most susceptible to wind erosion, about 18% of the land is subjected to wind erosion.

For Salinization

In world literature salinization is noted as a typical example of soil degradation, but precise information is not available regarding rate and hazard of this process. Even the classical examples referred to in the literature, such as the Tigris and Euphrates valleys, the Indus valley in Pakistan, the Nile and Senegal delta, the galodnaya steppe in USSR, etc., do not allow precise assessment of the salinization. The reasons for this are numerous, but fundamentally they are the dynamism of the process and man's efforts to improve saline lands.

Many models exist for assessment of salinization hazard. Most of them include the following elements:

-Ratio precipitation/potential evapotranspiration. It must be noted that one and the same ratio can have different effects on salts accumulation in the soil. Thus, for example, in the sub-tropical zone the ratio P/ETP = 0,4 is observed in regions with 250 mm precipitation, where saline soils occur widely. In contrast, with the same ratio in the tropical zone, precipitation is 700 mm and no saline soils are observed. This shows that the ratio P/ETP must be used differentially for each zone.

-Maximum accumulation of salts in the soils in the various bio-climatic zones.

-Capillary rise of the water, depending on the soil texture.

Table 1 - Ratio of Texture to Land Slope

Textural group of soil mapping unit	Percentage of soil mapping unit with			World land surface in	
	coarse texture	medium texture	fine texture	Sq. Km. x 1000	%
coarse textured	≥50	-	-	23495	17,3
medium textured	-	≥50	-	83052	61,2
fine textured	-	-	≥50	27908	20,6
equilibrated	<50	<50	<50	1220	0,9
Slope group of soil mapping unit	**Percentage of soil mapping unit with**			**World land surface in**	
	0-8% slope	8-30% slope	>30% slope	Sq. Km. x 1000	%
level to gently undulating	>60	-	<20	61600	45,4
rolling to hilly	<80	0-100	<50	53942	39,8
steeply dissected to mountainous	-	-	>50	20133	14,8

-Depth and quality of ground water.

-Topography of the region and status of soil cover.

-Additional elements, such as mineralization of surface flowing water, high and low tides, composition of the rocks in the water catchment basin, microrelief, type of salinization, irrigation methods, etc.

Inclusion of these elements (or a part of them) in models can provide a most general orientation about salinization hazard. These calculations, however, have a value only in an overall scale, where zones of relatively greater hazard can be indicated. Specific regions, however, require a differentiated approach for each of the indicated elements. The calculations made for Africa show that regions with zero to slight hazard have an index lower than 3,2, while those with very great hazard have an index greater than 39,0.

According to recent calculations, some 7,1% of the world's land has soils affected by salinization to various degrees. Based on the soil-climatic condition, it can be assumed that some more 8,2% additional areas actually are subject to this process.

Soil Status

The FAO/UNEP methods (1983) treat the soil status separately for each process. For an overall assessment, however, this is why other elements are taken into consideration. These elements include:

- soil requirements of the main crops towards the soils;
- soil irrigability;
- diagnostic horizons and soil properties;
- variation of the soil cover;
- land suitability (in three classes) for each one of the main crops.

The soil constraints index obtained as a result of combinations of these elements provide information about unfavorable properties for plant growth. The index is used for general assessment of desertification hazards. Soils with an index value of 4,0 or less are considered very good, and those with an index value of 20 or more are considered very poor.

Accumulation of toxic substance for plants, animals, and men has a local specifity and cannot be treated on an equal level with the above categories. An assessment of this process requires specific studies for each region affected by pollution.

Table 2 - Results for Africa

Rating Class	Soil Constraints	Water Erosion	Wind Erosion	Salinization	Population Pressure	Animal Pressure
	Surface (in %) of Africa by rating class and process					
None to slight	4,5	84,3	68,1	74,8	54,8	39,2
moderate	40,4	12,6	22,8	7,8	31,9	42,4
severe	42,8	2,1	3,8	6,6	10,6	10,4
very severe	12,3	1,0	5,3	10,8	2,7	8,0

Table 3 - Results for USSR

	Degradation of vegetation cover	Water Erosion	Salinization	Water erosion	Bad land (soil status)	Areas not prone to desertification
km^2	770083	59036	132081	12228	58584	650483
%	43,2	3,3	13,0	0,7	3,3	36,5

Desertification Hazards

The data obtained for each of the above categories (Anthropogenic Effect, Land Vulnerability, and Soil Status) serve to establish an index of the overall desertification hazards. Desertification hazards are small when the overall index is lower than 42,2. When this index is greater than 127,5 the hazards are very great. By calculation, the degrees of the overall desertification hazards are presented separately for regions with the following climatic characteristics:

- Areas without growing period, the most part of which are real deserts and only small areas (oases) can be subject of desertification;
- Areas with 1 to 180 days of growing period, which are most sensitive to desertification processes;
- Areas with more than 180 days of growing period and mountain territories with low temperatures, which are not subject to desertification.

Results for Africa and USSR

Due to the shortage of information for assessment of desertification in Africa, the interpretation method was applied (Boyadgiev T.). The results are as shown in Table 2 (FAO/UNEP/ESRI).

These data show that more than 95% of Africa has soils with unfavorable properties for one or more crops and that the negative anthropogenic influence can lead to quick land degradation. In addition, animal pressure (on some 60% of the territory) and population pressure (on some 45%) have a moderate to very great effect. From 15 to 32% of Africa is vulnerable to wind action, salinization, and water action.

Desertification hazards of the USSR Arid Lands (Babaev, A.G., I988) appear in Table 3. These data show that the processes leading to desertification of the USSR arid lands are mainly degradation of vegetation cover and salinization. The effect of wind erosion is low and that of water erosion is insignificant. The surface of the bad land is reduced.

Each one of these processes is manifested with different force in the individual countries. Specific measures for restricting, stopping, and/or changing the trend of this process must be planned, depending on the major degradation process.

References

Babaev, A.G.(1988) Explanatory Note to the Map on Manmade Desertification of the USSR Arid Lands, Scale 1 : 2,500,000. Ashkhabad, Turkmen SSR Academy of Science.

Boyadgiev, T.G. (1984) Report on the Modelling for compilation of the Maps of Desertification Hazards in Africa and soil elements used in assessing Desertification and Degradation in the world, FAO, Rome.

FAO/UNEP (1983) Provisional Methodology for Assessment and Mapping of Desertification, FAO, Rome.

FAO/UNEP/ESRI (1984) Map of Desertification Hazards, Explanatory Note.

Le Houerou, H. N., and F.H. Hoste. (1977) Rangeland Production and Annual Rainfall Relations in the African Sahels - Sudanian Zone. Journal of Range Management, Vol. 30, No. 3, Adsdis Ababa.

II. Soil Formation in the Arid Regions of Israel*

J. Dan[1]

Abstract

Three climatic subregions i.e. the extremely arid zone, the arid semidesert zone and the mildly arid zone, are defined in southern Isreal. These climatic zones are crossed by four physiographic subregions that include mountainous areas, large valleys, plains and sandy areas. All subregions are characterized by a certain soil and vegetation pattern.

The mountainslopes on hard rocks in the extremely arid zone are bare of soil cover. On soft rocks some physical weathering may be found. Reg soil formation characterizes mountain plateaus. The valleys are characterized by accumulation of eroded material that ranges from stones and gravel in the alluvial fans to fine sand, silt and clays in the inland depressions (playas). Both dry and wet saline playas are found in these areas.

Reg soil (Gypsiorthids) that are characterized by a desert pavement cover the large plains. Coarse desert Alluvium (Torrifluvents) and some fine desert Alluvial soils are found in wadibeds. Seif dunes characterize the desert sandy areas. Fine desert alluvial soils, mixed with alluvial sand, are found along ancient and recent wadis that cross these dunes.

Brown Lithosols (typic Torriorthents), both saline and nonsaline, are found in rocky pockets among hard rocks on mountain slopes in the arid, semidesert zone. On chalks and marls they are replaced by saline calcareous Lithosols and rendzinic desert Lithosols.

Young loessial sediments (Torrifluvents, some Xerofluvents) are found in valleys among these mountains. On terraces, plains and mountain plateaus, especially in the moister part of this region, loessial Serozems (Haplargids) are found.

Calcic and petrocalcic horizons characterize older gravelly and sandy sediments. Sand fields are found on recent sand.

Alluvial Brown soils and alluvial silty-clayey Serozems are found on alluvial fan material in the lower Jordan Valley. Saline and gypsiferous highly calcareous Serozems were formed from Lacustrine sediments on the Lissan terrace while marly saline desert Lithosols are found on badlands that dissect this terrace. Brown Alluvial soils cover most of the Jordan floodplain. Solonchacks are found in depressions and the lowest part of the Jordan floodplain.

Nonsaline brown Lithosols (lithic Torriorthents) are found on mountainslopes in the mildly arid parts of the northern Negev. Towards the north and moister parts they merge with Brown Rendzinas (lithic Xerorthents) and even with Terra-rossas.

Loessial soils, mainly loessial Brown soils (calcic Haploxeralfs), cover footslopes and small valleys in the northern Negev. Towards the north they are replaced by natric-grumic Serozems (Natrargids) and finally by natric Grumusols and reddish brown Grummusols (Typic Chromoxererts).

Loessial Brown soils (calcic Haploxeralfs) usually covering clayey paleosols up to 12 meters, cover most of the plains of the northwestern Negev. Young loess deposits are found in depressions.

Sandy Regosols characterize the recent sandy deposits of the western Negev. At depth they cover mature paleosols. Sand dunes are found near the coast.

Brown and reddish brown Grumusols (Chromoxererts) were formed from alluvial material on the Lissan terrace in the northern part of the Jordan valley. Hydromorphic Grumusols (Pelloxererts) cover impeded drained areas. Highly calcereous inseptic Brown soils (calcixerllic Xerochrepts) are found on lacustrine sediments on the Lissan terrace while rendzinic desert Lithosols and pale Rendzinas cover the terrace escarpment. Brown alluvial soils are found on the Jordan floodplain.

Introduction

Most arid regions, especially those of Israel, are heterogeneous in many respects. The mean precipitation ranges from about 350 mm down to less than 50 mm, and consequently the climate varies from mildly arid to extremely arid (Rosenan, 1970) (Fig. II.1). Those differences in climate affect the vegetation, landscape, and soil formation and, as a result, also the land use. In order to emphasize these interrelationships, three climatic subregions have been defined (Dan and Raz, 1970).

Climatic Subregions

The first is the extremely arid zone, where the annual precipitation is usually less then 80 mm. The vegetation in this zone is restricted to favorable ecological sites like dry riverbeds and rock

[1]Inst. of Soils and Water, Agricultural Research Organization, The Volcani Center, Bet Dagan, Israel.

*Published in 1981 as part of the field-guide of the Arid Soil Conference tour (Aridic Soils of Israel, Spec. Publ. No. 190, Agr. Res. Org., Bet Dagan)

Fig. II 1. Climate zone map.

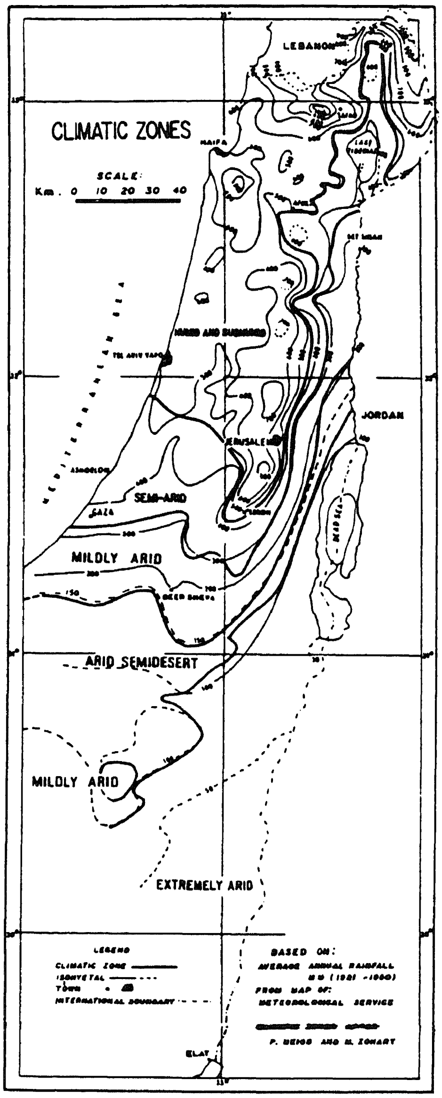


Fig. II 2. A schematic cross section of soils in the extremely arid zone.

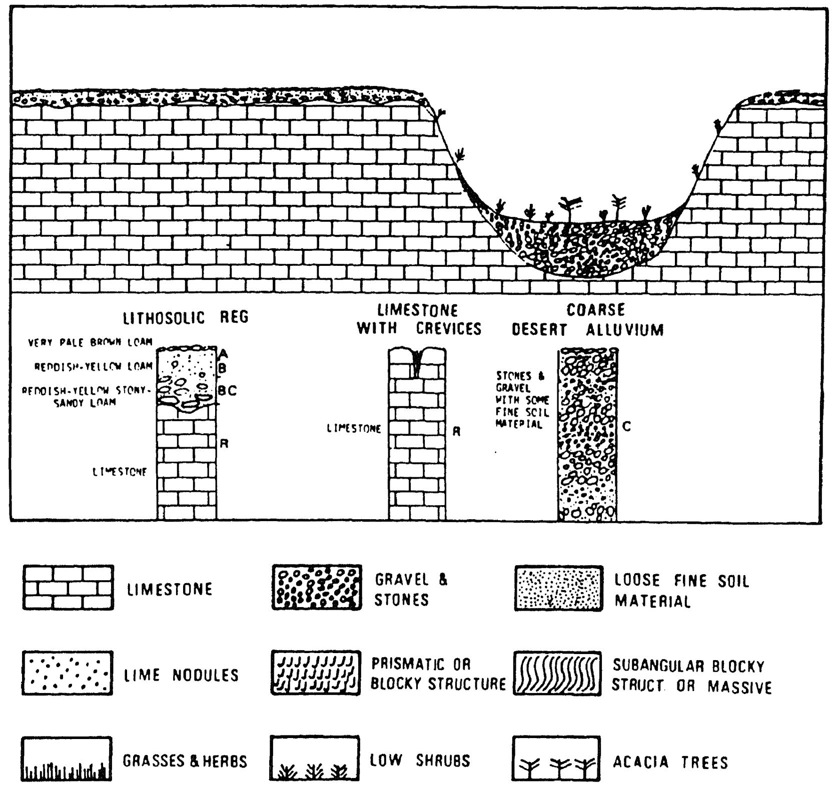
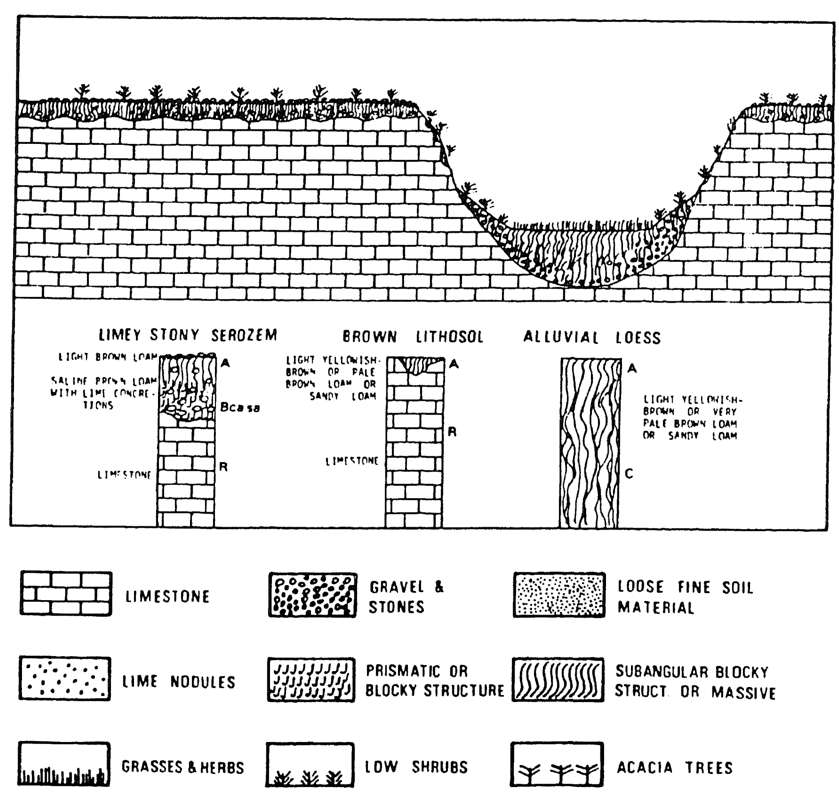


Fig. II 3. A schematic cross section of soils in the arid semi-desert zone.



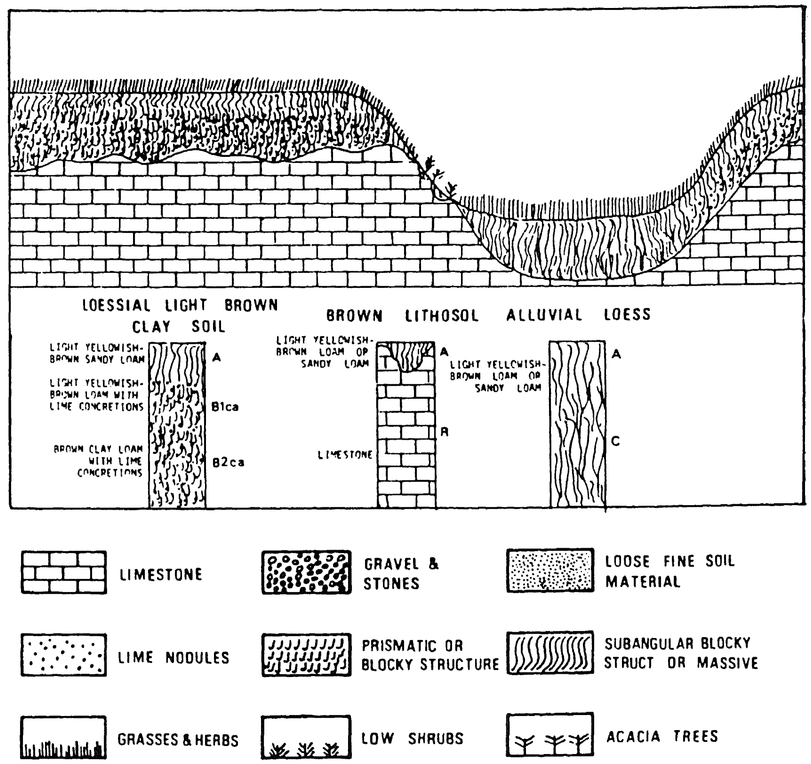
crevices (Zohary, 1955; Waisel et al., 1978) (Fig. II.2). Due to the absence of vegetational cover, erosion - both fluvial and aeolian - is severe. Hence, the surface deposits are depleted of fine soil material which is redeposited in the moister parts of the country (Yaalon and Dan, 1974). Soil development is very slow because of the extreme dryness. On old stable desert surfaces, accumulation of airborne salts becomes marked.

The second zone, the arid semi-desert zone, is located wherever the annual precipitation ranges from 80 mm to about 200 mm. The vegetation of this zone is diffuse (Danin et al., 1975) and cannot prevent severe

erosion. In favorable ecological sites like footslopes or depressions, however, where the runoff water concentrates, the vegetation is more dense and favors deposition of the finely or eroded material (Yaalon and Dan, 1974)(Fig. II.3). This material consists mainly of loess that had been deposited on the hillslopes and was re-eroded due to the absence of a protective dense vegetation cover. Soil development is usually slow, due to the dry climate, but is somewhat more marked than in the drier, extremely arid zone.

The third zone is the mildly arid climatic zone, where the annual precipitation ranges from about 150 or 200 mm to about 350 mm. This amount of rainfall usually enables the formation of a continuous vegetational carpet (Fig. II.4), and most of these areas have been characterized in the past by a low steppe. The vegetation restricts soil erosion; as a result, finely or weathered material covers the plains and moderate slopes. The protective cover of this vegetation also enables the heavy accumulation of Loess in this region (Yaalon and Dan, 1974), as it prevents re-erosion of this material. This region receives heavy loads of fine aeolian dust that has been eroded from the drier desert regions (Yaalon and Ganor, 1975). Soil profile differentiation processes are significant, and mature, well differentiated soils are quite widespread.

Fig. II 4. A schematic cross section of soils in the mildly arid part of the northern Negev.



Physiographic Units

Several broad physiographic units can be defined in the arid areas of Israel. Part of the land is mountainous, while other parts consist of plains, plateaus, or undulating areas (Fig. II.5). Large valleys, especially the large, young rift valleys, form another broad physiographic unit. Finally, the areas of sand dunes in the desert represent a separate unit, due to their specific features.

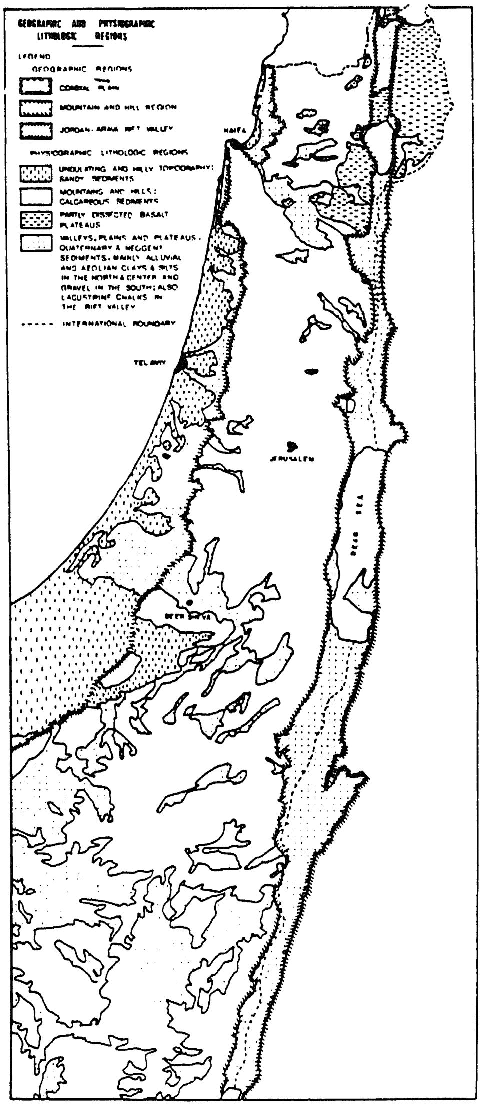
The landscape, soil formation and pattern, and vegetation pattern, etc., are quite different in each of these physiographic units, even within the same climatic zone (Dan and Raz, 1970). Erosional processes are severe in the mountainous areas, while depositional processes characterize the young rift valleys. In the center of these valleys, saline marshes and saline lakes usually are found, because wadi channels do not reach the sea. In the plains, both erosional and depositional processes are minimal; as a result, soil formation processes are more marked. In the sand areas, however, wind erosion and shifting sand inhibit soil formation.

Each of the aforementioned physiographic units can be found in the three arid climatic zones. It was thus thought desirable to define climatic-physiographic units, each of which is characterized by a unique landscape, soil, and vegetation pattern (Dan, 1979a). In the following pages the soil formation of each of these climatic physiographic units is described. Table II.1 provides a list of the soils of arid lands in Israel in relation to climate and physiography, and Table II.2 a description of the vegetation.

Soil Formation in the Mountainous Part of the Extremely Arid Desert in Israel

The weathering in the mountains is very slow, due to the dry climate. Shallow saline calcareous desert Lithosols[2] (Lithic Torriorthents) were formed on the soft carbonate rocks, like chalk and marl, due to the rapid mechanical weathering of these rocks. These Lithosols consist actually of the physically broken down rock mate-

Fig. II 5. Geographic and physigraphic-lithologic regions of Israel.



rial, so the chemical feature of these soils resembles that of the underlying bedrock (Dan and Raz, 1970). These Lithosols are very saline, and it seems that this salinity is higher than that of the underlying rocks.

This is the result of some salt leaching of the uppermost soil layers and the concentration of the salts beneath the soil surface. The uppermost soil layer is later eroded, but some of the salts of this layer remain in the soil.

[2]Great group soil names according to the classification of Israel soils are written with a capital letter (Committee on Soil Classification in Israel, 1979). They are mostly correlated with the U.S. Classification (Soil Survey Staff, 1975).

The hard rocks on the slopes are usually bare of soil cover, due to their slow breakdown and the severe erosion. On plateaus some shallow Reg soils (Typic Gypsiorthids and Petrogypsic Gypsiorthids) were formed on hard rocks (Dan, 1979b), a feature which characterizes mainly the areas where flinty strata are common. In such cases the flint gravel accumulates on the surface and protects the soil from accelerated erosion. Soil formation in these places is similar to the development of the Regs on the plain and is described in a later section.

The erosion in the mountains is severe due to the climate, steep slopes, and absence of vegetational cover. As a result, most of the weathering products are carried away. The coarse alluvial material, which contains many stones and gravel (Typic Torriorthent or Typic Torrifluvent), is deposited on footslopes, in small valleys, in the streambeds, and along the alluvial fans which cover large areas in the valleys between the mountains.

This coarse material in the small valleys is usually young and unweathered. The soils that are found in these valleys are thus defined, according to the nature of the sediment, as coarse Desert Alluvium if they contain mainly stones and gravel, or as stony and gravelly sandy Desert Alluvium if they contain also large amounts of sand (Committee on Soil Classification in Israel, 1979). In some elevated spots like terraces or stable dissected alluvial fans and cones, some young Reg soils (Typic Gypsiorthids) may be found. The formation of these soils is similar to that in the plains and is described in a later section.

Table II.1: Soils of Arid Lands in Israel in Relation to Climate and Physiography

Climate / Physiographic unit	Extremely arid	Arid semi-desert	Mildly arid
Mountains	Bare rocky slopes; Lithosolic Regs (on mountain plateaus). Saline calcareous desert Lithosols on chalks and marls; in gullies, coarse Desert Alluvium and stony and gravelly Desert Alluvium.	Brown Lithosols, usually saline. Saline calcareous desert Lithosols. In gullies, Loess and coarse Desert Alluvium.	Brown Lithosols, Brown Rendzinas, Rendzinic desert Lithosols; various shallow Brown soils, some Natric Grumusols. In gullies, Loessial stony Brown soils and various colluvial-Alluvial soils.
Great valleys (usually rift valleys)	Coarse Desert Alluvium. Gravelly loamy Alluvium. Stony - sandy Alluvium and gravelly sand. Aeoliam and Alluvial sand. Fine desert Alluvial soils; Takyrs. Alluvial Solonchaks, Alluvial sterile Solonchaks.	Various Serozems (including argillic Serozems), mainly loessial and stony Serozems (in Negev mountains) or alluvial silty-clayey Serozems (in the Jordan Valley). Loess and stony alluvium (in Negev mountains) and various calcareous Alluvial soils (Jordan Valley).	Grumusols, highly calcareous Brown soils and brown silty Alluvial soils (in the Jordan Valley). Loessial and stony Brown soils and Alluvial Loess (in the Negev).
Plains, plateaus and undulating areas	Regs. In depressions, coarse Desert Alluvium and some Loamy Desert Alluvial soils.	Stony Serozems; Alluvial Loess in depressions. Loessial Serozems in transition to the midly arid zone.	Loessial light Brown soils and some other light Brown soils. Alluvial Loess in depressions.
Sand plains	Sand dunes, some Sand fields, some fine Desert Alluvial soils.	Sand dunes, Sand fields.	Sandy Regosols, Sand fields, some sand dunes.

Soil Formation in the Large Desert Valleys

Most of the eroded material from the mountains is deposited in the large desert valleys. The coarse stony and gravelly sediments were deposited usually along the alluvial fans, while the fine material - especially fine sand, silt, and clays - was carried farther away toward the depression or to the lakes and the sea (Dan, 1979a, b). The soil formation on the alluvial fans is usually negligible, while in the depressions it is usually more pronounced due to the special hydrology of these areas. The soil formation of these two geographic facets will be described separately.

Table II.2: Vegetation of Arid Lands in Israel in Relation to Climate and Physiography

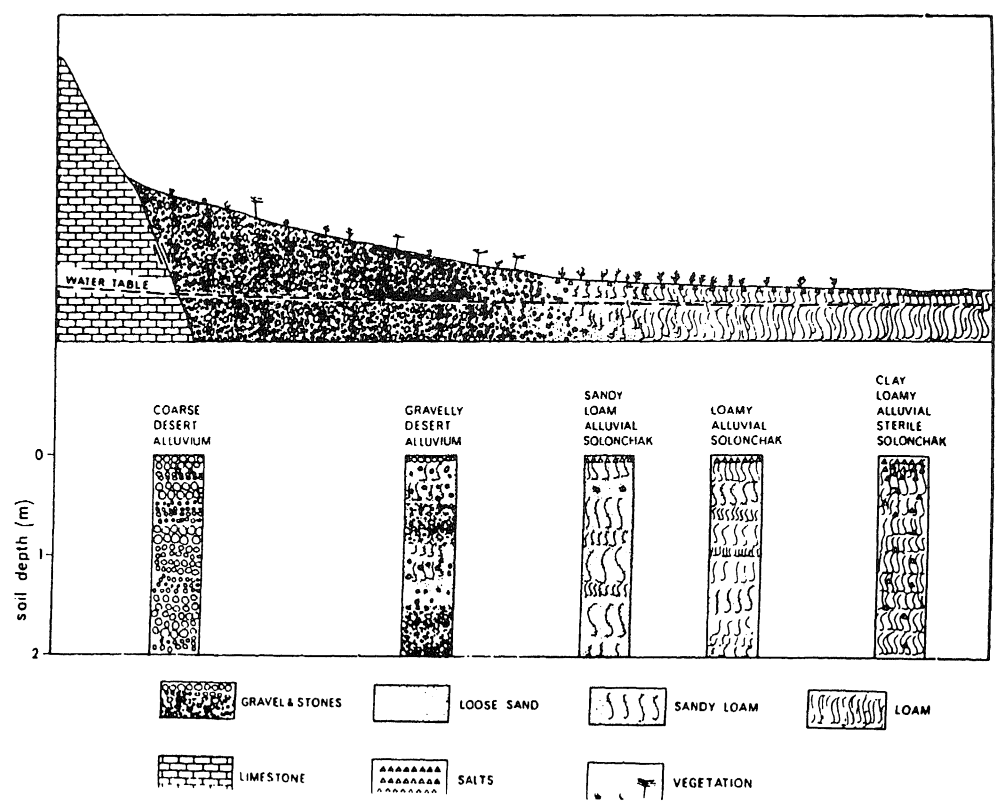
Physiographic unit \ Climate	Extremely arid	Arid semi-desert	Mildly arid
Mountains	Scattered low desert shrubs, mainly of various Zygophyllum dumosum plant associations on mountain on hard rocks. Low desert shrubs and some Acacia trees in rivulets.	Desert shrubs, mainly of various Artemisia herba-alba plant associations on mountain slopes; in dry parts also Zygophyllum dumosum plant associations; quite dense shrub and grass vegetation at valley bottoms.	Desert shrubs mainly of various Artemisia herba-alba plant associations; in many parts also Sarcopoterium spinosum plant associations
Great valleys (usually rift valleys)	Low desert shrubs mainly of Anabasis articulata plant associations accompained by acacia trees in stream-beds; in sandy areas scattered Haloxylon persicum shrubs. Halphytic shrubs on Solonchaks.	Scattered low desert shrubs, mainly of Hammada scoparia and Anabasis syriaca plant associations (in the Negev); Halophytic shrubs in Solonchaks (Jordan Valley).	Steppe vegetation of grasses and herbs.
Plains, plateaus and undulating areas	Low desert shrubs mainly of Anabasis articulata associations accompained sometimes by acacia trees in streambeds; no vegetation on Reg soils.	Scattered low desert shrubs, mainly of Hammada scoparia plant associations on plains. Quite dense shrub and grass vegetation in streambeds.	Steppe vegetation of grasses and herbs
Sand plains	Scattered desert shrubs, mainly of Haloxylon persicum.	Quite dense cover of desert shrubs, mainly of various Artemisia monosperma plant	Steppe vegetation of grasses and herbs.

Soil Formation along Alluvial Fans

Most of the coarse material which was eroded from the mountains that border the Arava Valley (the rift valley south of the Dead Sea) has been deposited in the alluvial fans. These alluvial fans cover large parts of the valley, and especially the southern parts (Ron, 1967). The coarse debris, which includes mainly large stones, has been deposited in the upper part of the fans, while the somewhat smaller-sized sediment, mainly gravel and sand, has been deposited in the lower part of the fans. The still finer sediment, like fine sand and clay, has been carried farther on to the playas and sometimes even to the sea.

The composition of the alluvial fan sediments is related to the rock types which were found in

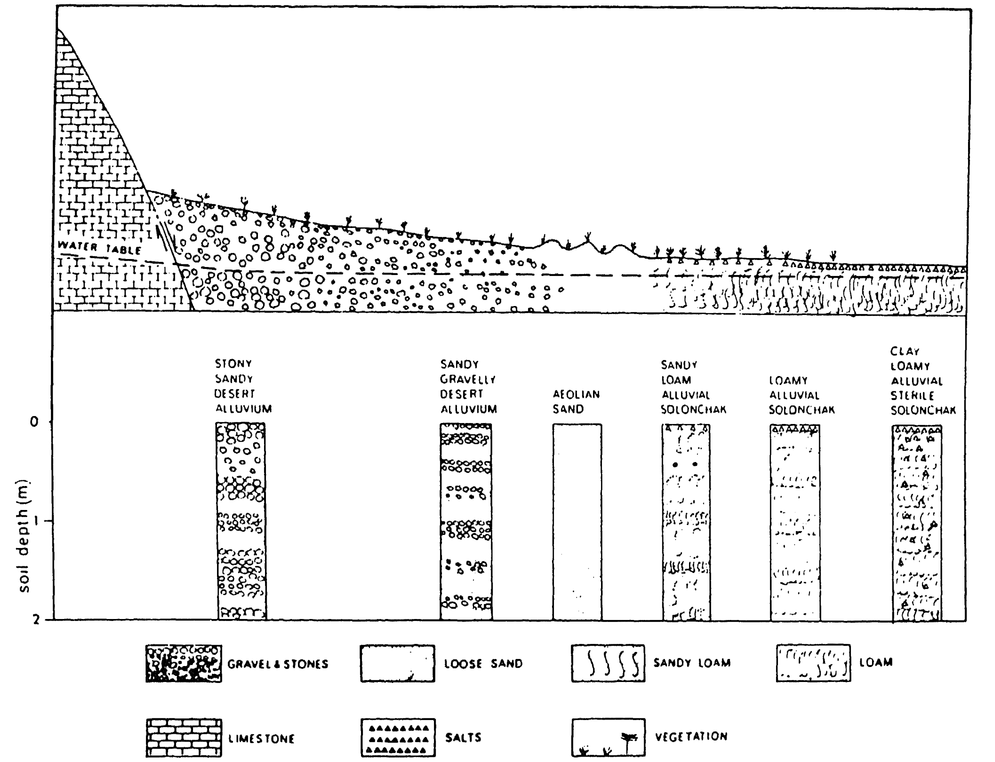
Fig. II 6. A schematic cross section along an alluvial fan and nearby Solonchak depression that formed from material which had been eroded from carbonate mountains.



the catchment areas (Dan, 1979a,b). These include mainly various carbonate rocks and flint strata (Picard, 1970). Such rocks do not form much sand during the mechanical weathering processes; as a result, the sediments along most of the alluvial fans in Israel range usually from coarse Desert Alluvium (Typic Torriorthents or Typic Torrifluvents) in the upper parts of the fans to gravelly loamy Alluvium (Typic Torriorthents or Typic Torrifluvents) in the lowest parts of these fans (Fig. II.6).

In the mountain bordering the southern part of the Arava, sandstone and granite rocks are also quite widespread (Picard, 1970; Ron, 1967). In this case the fan sediments from these mountains contain much sand. As a result, a sandy belt may be recognized in the lower part of the alluvial fan (Dan, 1979a,b). Thus, the sediments on these alluvial fans range from coarse Desert Alluvium (as above) in the uppermost parts of the fans, through stony sandy Alluvium (Typic Torriorthents or Typic Torrifluvents) in the middle part, to gravelly sands (Typic Torripsamments) in the lower parts of the fans (Fig. II.7). Some of the sand, in the lower part of the fan, may be picked up by wind, due to the absence of gravel pavement, and form some small dunes (Typic Torripsamments) along the lowest part of these fans.

Fig. II 7. A schematic cross section along an alluvial fan and nearby Solonchak depression that formed from material which had been eroded from mountains where sandstone rock are widespread.



Sedimentation on the alluvial fans continues intermittently also today, and sediments do not reveal, as a rule, any soil formation (Dan and Raz, 1970; Dan, 1979b). In inactive parts of the fans, some very shallow Reg soil formation may be seen.

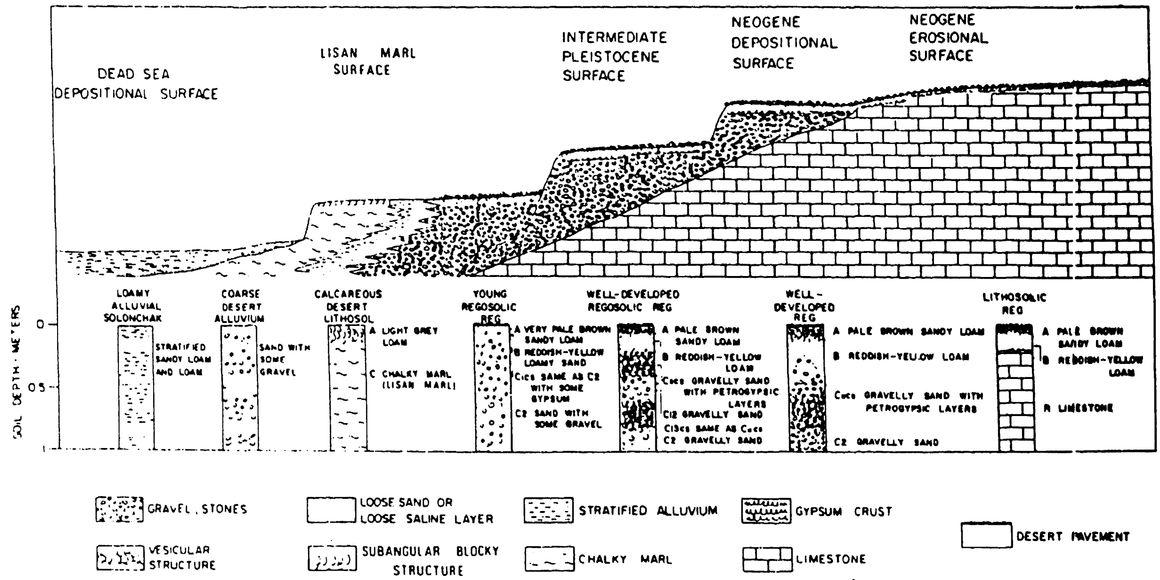
Soil Formation in Undrained Depressions

Most of the fine alluvial sediments, (Typic Torrifluvents) i.e., fine sand, silt, and clay, are deposited only when the carrying capacity of floodwater is reduced to very low values. This occurs only in the depressions among the alluvial fans or in very broad dry riverbeds with a negligible gradient, like Wadi el Arish. The water table in some of these depressions is found at a shallow depth. This feature characterizes mainly the depressions near the seacoast, like that of Sodom and Elat, or other depressions which were found at low elevation, like that of Yotvata and Avrona (Amiel and Friedman, 1971; Dan, 1979a,b).

Soil Formation in Wet Playas

The groundwater in these depressions is saline. The salinity *increases toward the center of the depressions*. The soils in these depressions, especially their upper layers, are salinized due to capillary rise and evaporation of groundwater and occasional floodwater. In peripheral areas where the water table exceeds 2 meters, the salinization may be caused by deep-rooted plants that absorb the salts with the water from the deeper soil layers and return them to the soil surface via the leaves and the plant litter, as occurs with the *Tamarix* plant (Waisel et al., 1978). As a result of this kind of salinization, some typical Solonchaks (Salorthid also some Typic Torrifluvents) soils were formed.

Fig. II 8. A schematic cross section revealing the relationship between soil profile characteristics and the various surfaces in the Central Arava Valley.



Soil texture becomes finer toward the center of the depressions, as the finer particles are carried further away during the floods. As a result the soil grades from sandy and sandy loam alluvial Solonchaks (Salorthids also Typic Torrifluvents) in the periphery of the depressions to loamy and clay loamy Solonchaks closer to its center. The center of the depressions is usually occupied by fine-textured alluvial Sterile Solonchaks (Salorthids) which are bare of vegetation due to the high salinity of both the soils and the groundwater.

In some places, where the water table is very high for most of the year, the various soil layers, and especially the deeper ones, suffer severely from impeded drainage. As a result, reduction processes occur and Gleys (Typic Salorthids) are formed. These Gleys are found usually in the central part of the depression, where fine-textured gley Solonchaks, and especially fine-textured sterile gley Solonchaks occur. However, gley Solonchaks (Typic Salorthids) may be found also in peripheral areas, between the alluvial fans where the land surface is much lower than in neighboring fan areas.

Springs are usually present in these areas. The vegetation of these peripheral areas is frequently quite dense as the groundwater is not strongly salinized and, as a result, alluvial gley Solonchaks occur in these places. Some of these soils may also contain quite a high content of organic material due to the dense vegetation and the impeded drainage conditions; in these cases Aquollic Salorthids are recognized.

Soil Formation in Dry Playas

Soil formation is usually negligible in depressions where the water table is deep (Dan, 1979b). The same may hold true for the soils of Wadi el Arish and its tributaries. Thus, these soils have characteristics resembling those of the alluvial sediment and were designated as fine Desert Alluvial soils (Committee on Soil Classification in Israel, 1979) (Typic Torriorthent or Typic Torrifluvents). However, in some of these depressions, like Qa en Naqb, the soils have been severely salinized due to the concentration of salts in the ponded floodwater. As a result, the sodium concentration in the exchange complex reached high values. The ponded floodwater had, to some extent, leached the uppermost soil layer and, as a result, a hard sodic crust was formed in these soils. These soils have been classified as Takyrs, like similar soils in Central Asia (Kovda, 1973).

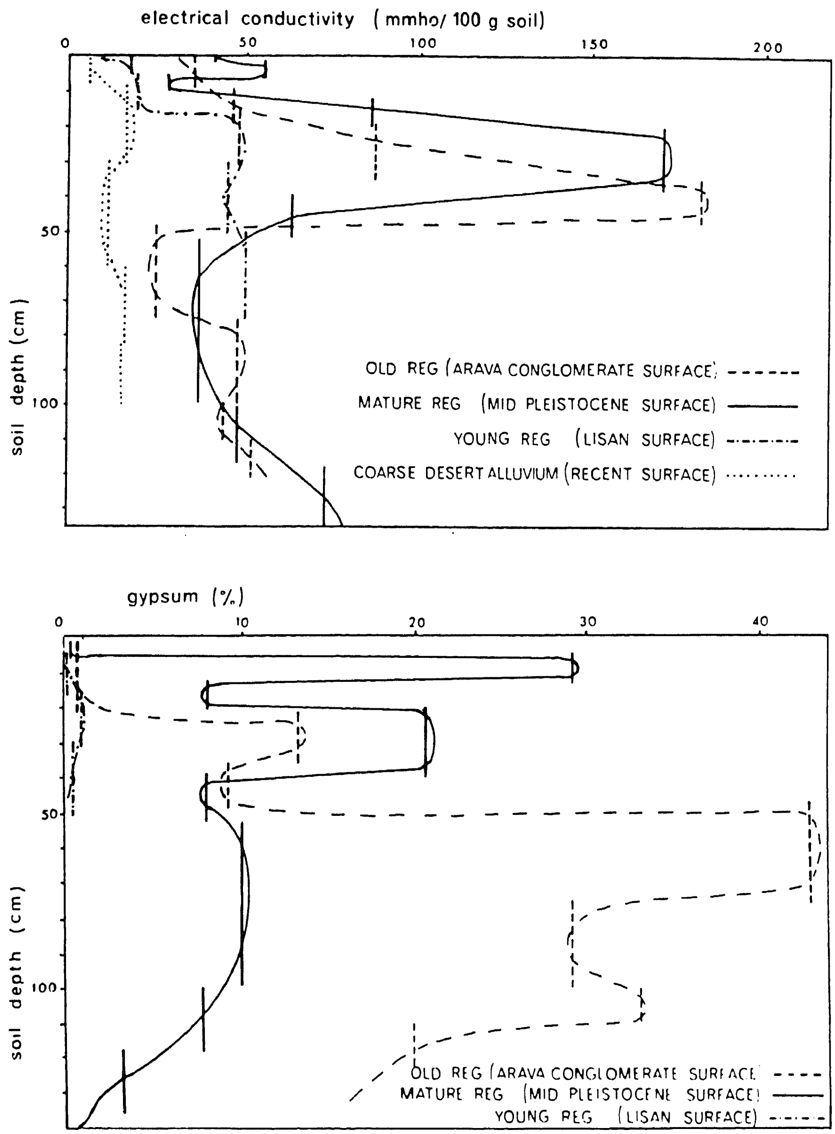
Soil Formation on the Gravel Plains

The gravel plains were formed from old desert valley fill which was composed mainly of coarse alluvium (Horowitz, 1979). The sediments were subjected to severe water and wind erosion that carried away all the fine material from the surface; as a result, a cover of gravel and stones resistant to weathering was gradually formed on the soil surface (Evenari et al., 1971; Dan, 1979b).

Erosion decreases to very low values after formation of the gravel cover, due to the protective

feature of the gravel and stones. As a result, soil profile differentiation and soil formation processes affect these soils. These are very slow, due to the dry climate; they include mainly atmospheric salinization and concentration of gypsum from airborne salts (Yaalon, 1963), but some clay formation or introduction can also be recognized. The soils which were formed due to the above-mentioned combination of processes are defined as Regs (usually Gypsiorthids).

Fig. II 9. Salinity and gypsum distribution in Reg profiles from various surfaces.



Stages of Reg Soil Formation

Several stages of Reg soil formation can be recognized. On quite young sedimentary surfaces, like the Lisan surface near Hazeva and inactive parts of alluvial fans, soil development is restricted mainly to the formation of a thin, 1-2 cm, vesicular layer and an underlying thin, somewhat reddish layer (Fig. II.8). Some salt and gypsum concentration in the deeper layers characterizes these soils (Fig. II.9). This soil will be still included among the Torriorthents.

The various soil layers become thicker in the older geographic surfaces (see Fig. II. 8). This development is accompanied by severe salinization and concentration of gypsum (see Fig. II.9) The gypsum usually forms several hard petrogypsic horizons in the deeper soil layers. Quite deep Reg soils with several petrogypsic horizons (Petrogypsic Gypsiorthids) characterize most of the large gravel plains in the Paran Desert and central Sinai.

The Typical Reg Soil Profile

The typical mature Reg soil (Petrogypsic Gypsiorthid) consists of several horizons (Dan et al., 1982; Dan, 1979a). Underneath the desert pavement a very pale brown loamy vesicular horizon of several centimeters is found. The next horizon, of about 10 to 20 cm, consists usually of a reddish-yellow dusty and very saline loam or clay loam. Soft gypsum or anhydrite chunks are usually found in this horizon.

The deeper layers usually consist of a somewhat mechanically weathered material. The fresh parent material, consisting of a mixture of gravel and stones with some finer materials in regosolic Regs, or of various carbonate rocks in lithosolic Regs, is found usually at a depth of about 60 to 90 cm. Hard gypsum crusts - petrogypsic horizons - are found usually in the deeper layers of these soils. The Reg soils are bare of vegetation.

Other Soils in the Reg Plains

Some of the Reg plains are dissected due to the drop of the base level. The gravel and stones

that are found on the dissected slopes do not reveal any soil differentiation and they may thus be defined as gravelly or stony Regosols (Typic Torriorthents). (Committee on Soil Classification in Israel, 1979).

The gravel plains are transversed by many dry water courses (wadis). Desert alluvium, especially coarse desert alluvium (Typic Torriorthents or Torrifluvents), has been deposited in these water courses (Dan, 1979a,b). The composition of this alluvium depends on the carrying capacity of the floodwater; in the large, main wadi the floods are, as a rule, heavy, and the streamflow is rapid, so that only coarse materials, like stones, are left behind. In the smaller wadis the stream is generally much slower, and as a result gravel and even sand and silt are deposited.

The streamflow in the central part of Wadi el Arish and its tributaries is generally very slow due to the very small gradient, and fine sediments - especially fine sand and silt - have been deposited in these wadis. A great part of this fine sand and silt, as well as similar material that was left on the gravelly wadi beds at the end of the flood, are carried away by winds and form the source of loess deposits (Yaalon, 1978).

Soil Formation in Sandy Deserts of Israel and Northern Sinai

Soil formation in sandy deserts of northern Sinai and the western Negev is negligible. The shifting sand is not, as a rule, stable enough for soil horizons to be formed. Even in sand fields (Typic Torripsamments) where the sand is more stable, no soil horizons are detected.

Most of the sandy region is covered by parallel shifting seif dunes (Tsoar, 1978). The seif dunes are oriented east-west as a result of the bi-directional seasonal winds.

Depressions between the dunes sometimes reach a width of 100 meters or more. The somewhat coarser lag sand in these depressions is usually quite stable and often covered by a rather dense vegetation, especially in the northern and eastern parts of the region. However, soil formation is still negligible (Dan, 1979a,b).

Several dry wadis reach the dune area from the south. Most of these water courses get lost in the dunes, since they do not carry enough floodwater to force their way across the seif dune barriers. The fine sandy and silty sediments carried by the floodwater are deposited between the dunes, which is why these valleys are covered by fine Desert Alluvial soils (Typic Torrifluvents). These soils are usually covered by some sand, and in many of them sandy layers are found among the finer deposits. The only dry rivers that now cross the dunes are Nahal haBesor and Wadi el Arish. Fine Desert Alluvial soils are also found here along the river in the recently formed flood plains.

The altitude of the sand dunes region ranges from sea level to about 300 m along the southern border of the region. As a result of the low altitude in the northwestern part of the region, groundwater is found close to the surface in the interdune areas (Dan, 1979a). This groundwater flows slowly from the south, northward toward the sea. The groundwater becomes highly salinized due to strong evapotranspiration, and Solonchaks (Salorthids) are formed as the saline groundwater rises to the surface. These Solonchaks cover large areas, especially near the sea and the lagoons.

In the southern part of the farther inland saline depressions where the groundwater is not very saline, there is usually some vegetation. Many date palms are grown in these places. Toward the sea coast, the vegetation gradually disappears, as a consequence of the high salinity of the groundwater; as a result, typical sterile Solonchaks (Salorthids) cover these areas.

Soil Formation in the Arid Semi-desert Parts of the Mountains in Israel (in the central Negev and the lower parts of the Judean Desert)

The weathering of the hard rocks in the Negev mountains is very slow, due to the dry climate. Thus, the soils of these mountains were formed mainly from the loessial deposits that have been mixed, to some extent, with the mechanical breakdown of the rocks.

These soils are confined to pockets and crevices among the rocks where they are protected from accelerated erosion. They are designated as Brown Lithosols (Lithic Torriorthents). Most of these soils are saline, due to the concentration of airborne salts, and, as a result, are defined as saline Brown Lithosols. However, these Lithosols are usually less saline than the deep soils in this region, as the rock pockets and crevices receive some runoff water from the rock exposures.

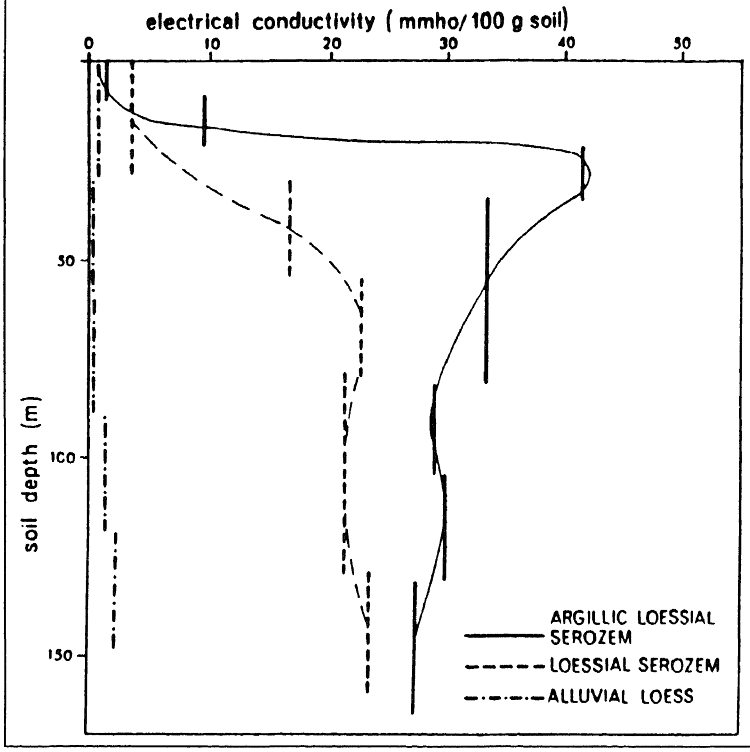
This concentration of local runoff water also enables the development of relatively rich vegetation (Danin et al., 1975). In very rocky areas where rock exposures cover most of the surface, the amount of local runoff increases, and in

some cases it may even leach the salts; thus, in these places, nonsaline Brown Lithosols can be found.

The mechanical breakdown of the soft rocks, like chalk or marl, is somewhat faster and, as a result, the soils resemble the composition of the underlying rocks. These soils are included among the calcareous desert Lithosols (Lithic Torriorthents). However, at somewhat moister sites, like northfacing slopes, especially in the moister part of the region, these soils may reveal some formation of an A horizon (Dan and Nissim, 1976). This is due to denser vegetation and, as a result, there is also some formation of organic material, which may form also a somewhat better structure. These soils have been defined as Rendzinic desert Lithosols (Typic or Lithic Torriorthents) due to their resemblance to the pale Rendzinas which are found in the moister parts of Israel.

The soils on the soft chalk and marls are more saline than those on hard rocks. In these places rock outcrops are usually absent so that the soil does not receive any local runoff water that can leach the soil. Moreover, runoff of the soil surface itself is severe, especially in the desert Lithosols, due to the absence of vegetation cover (Evenari et al., 1971) and the weak structure of the upper soil layer. As a result, these soils are severely salinized.

Fig. II 10. Salinity content of an Alluvial Loess, a Loessial Serozem, and a Loessial Argillic Serozem at Sede Boger. The differences in salinity are related mainly to soil age.



Soils of the Valleys and Plains in the Arid Semi-desert of the Negev and The Judean Desert

The mountains in the semi-desert areas are severely eroded, as no dense protective vegetation cover exists in this area (Yaalon and Dan, 1974). The eroded sediments contain, as a rule, loamy loessial materials which are sometimes mixed with gravel or stones. The gravelly and stony sediments are deposited usually along the footslopes and in the large dry wadibeds, while the loessial, stone-free sediments are depoisted along the small depressions and in the flooplains.

Soil Development from Loess

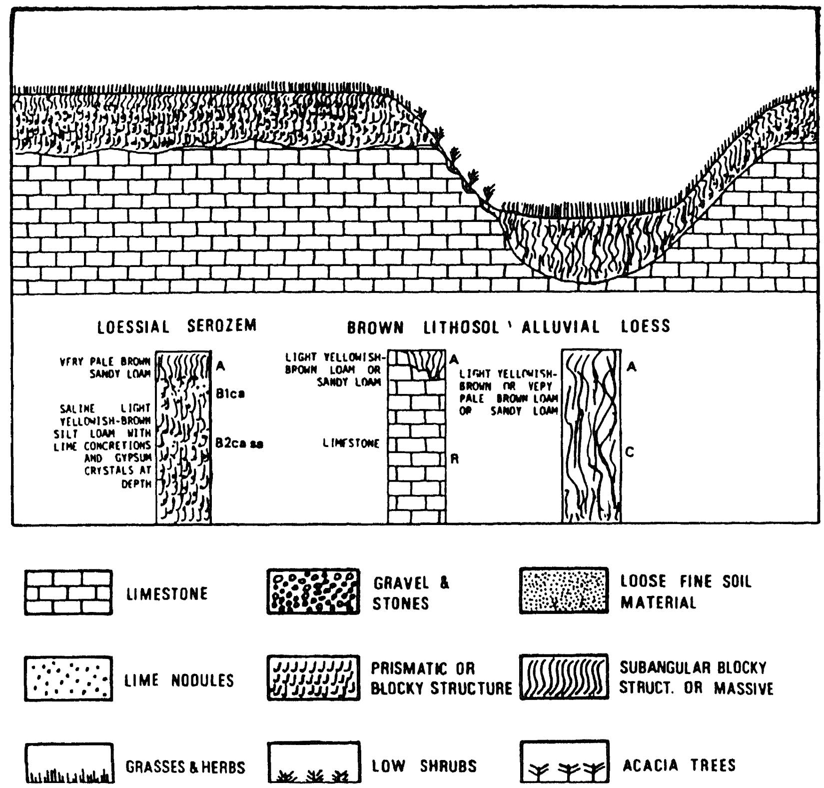
These young loessial deposits do not reveal any profile differentiation, as their deposition is rapid (Dan, 1979a); they are designated as Loess or alluvial Loess (Torrifluvents or Xerofluvents, according to the moisture regime). On the terraces, however, soil formation may already be revealed. Redeposition of lime at a shallow depth and slow salinization of the deeper soil layers caused by airborne salts characterize these soils (Dan, 1979a, b) (Fig. II. 10). Some clay illuviation may also be revealed.

These soils are defined as loessial Serozems. The B horizon of the young loessial Serozems is of a loamy cambic nature and it is designated as a loamy loessial Serozem (correlated with the Calciorthids) (Committee of Soil Classification in Isreal, 1979). With increasing development, usually in the higher terraces, the B horizon becomes somewhat finer textured and clay loam argillic loessial Serozems (correlated with the Typic Haplargids) are formed. Loessial Serozems also have been formed from aeolian loess. In the Negev mountains they are confined to flat palteaus where aeolian loess had been deposited during a presumably somewhat mositer period in the past (Dan et al., 1979).

Toward the northern edge of this area, aeolian loess has been deposited also on plateaus and northern slopes, due to somewhat more favorable ecological conditions (Yaalon and Dan, 1974). Loessial Serozems also have been developed from these deposits (Dan, 1979b), (Fig.

II.11). These soils are widespread in the Be'er Sheva' valley and the Arad plateau. The salinity of these Serozems decreases gradually toward the moister areas in the north until they merge with the non-saline loessial light Brown soils (Calcic Haploxeralfs, Calcic Palexeralfs or Calcixerollic Xerochrepts).

Fig. II 11. A schematic cross section of soils in the northern, moister part of the arid, semi-desert zone.



Soil Development from Gravelly and Stony Deposits

Soil formation on the gravelly and stony deposits resembles that on the loess. Young sediments in the dry riverbeds do not reveal any soil formation and are thus designated as coarse Alluvium (Torrifluvents or Xerofluvents). On terraces, footslopes, and inactive fans, clacic and cambic horizons develop and the soils are designated as gravelly and stony Serozems (Calciorthids) Fig. II.12). With increasing time an argillic horizon is sometimes formed (Argids according to the American classification).

Carbonate content in the calcic horizon of the gravelly and stony soil increases continuously with time until the lime coatings and nodules merge to form a petrocalcic horizon (Dan, 1977), and thus petrocalcic stony Serozems or petrocalcic Paleargids are formed. Old Neogene gravelly and stony sediments are widespread in the eastern Negev on high dissected terraces (Horowitz, 1979). Various stony Serozems, especially petrocalcic stony Serozems, cover most of these Neogene sediments.

Soil Development from Sand

A similar sequence of soil development can be seen in the sandy sediments that are widespread in some synclinal valleys in the eastern Negev and in the western part of the Negev. The primary stages of soil development include Sand fields and sandy Regosols (Typic Torripsamments). These soils are still widespread in the western Negev, where such areas have been covered by sand during the last few millenia. With the passage of time, sandy Serozems were formed.

The primary stage is characterized by a sandy A horizon and a shallow sandy cambic Bca horizon. This soil is designated as sandy quartzic Serozem (Camborthids or Calciorthids) (Committee on Soil Classification in Israel, 1979). With increasing time a loamy argillic Bca horizon is formed and the soil designated as loamy quartzic argillic Serozem (Typic Haplargids). The concentration of lime in the Bca horizon continues here, as in the stony Serozems, until a petrocalcic horizon is formed (Dan, 1977), and the soils are designated as petrocalcic quartzic Serozems (petrocalcic Paleargids).

These advanced stages of development are confined to sites of Neogene sand which are usually mixed with some stones, in the vicinity of Arad and Aroer. The carbonates of these soils originated mainly from airborne sediments (Dan, 1977), as the quartzic parent material does not usually contain carbonates.

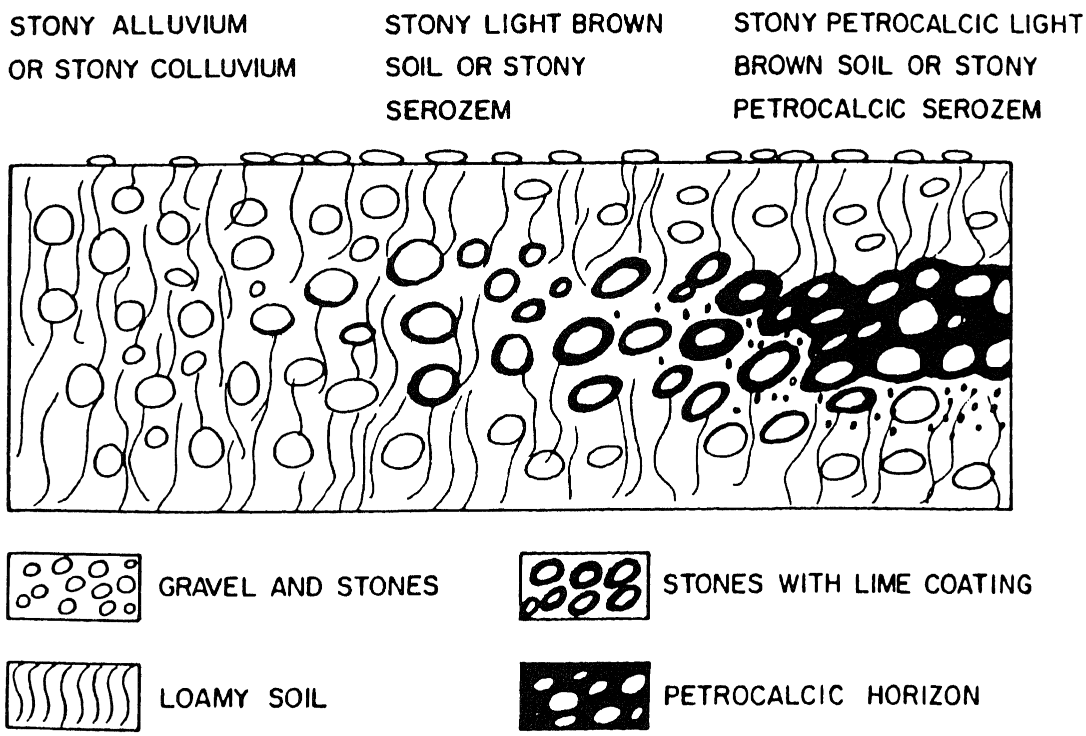
Soil Formation in the Southern Arid Part of the Jordan Valley

The soils of this region have been developed either from alluvial material - mainly along the alluvial fans - or from marly-lake deposits (Lisan marl) (Dan and Alperovitch, 1971, Dan et al., 1981).

The Soils of the Alluvial Fans

The large alluvial fans deposited clayey and silty soil material that had been eroded from soils of the more humid mountains. Various alluvial Brown soils have developed from this

Fig. II 12. Soil development from gravel in the arid semidesert zone and in the mildly arid part of the country (according to Dan, 1977).



material. Stony and gravelly alluvial Brown soils (Calcixerollic Xerochrepts) are found mainly on the upper parts of the fans; the soils become more fine-textured and more saline toward the lower parts of the fans and saline alluvial silty clayey Serozems (Typic Haplargids, some Vertic Haplargids) are formed.

It seems that these soils were salinized in the past from a high water table associated with the Lisan lake (the ancient Dead Sea). Subsequently, the soils were drained due to the dissection of the Lisan terrace, and nowadays they are affected mainly by the present dry-climate conditions. Some small alluvial fans that deposited highly calcareous material from the nearby desert are found among the large alluvial fans. Most of the soils in these fans do not reveal any profile differentiation and thus they are designated as highly calcareous pale brown Alluvial soils (Typic Torrifluvents or Typic Torriorthents).

The water table in the alluvial fans is deep, but it becomes shallow toward the lower part of these fans. The water is also very shallow in narrow strips between the fans; there groundwater caused severe salinization and as a result Solonchaks (Salorthids) are found along the strips. It seems that in the past, before the dissection of the Lisan terrace, or even during the existence of the Lisan lake, the extent of the areas which suffered from a high water table was larger, and it is possible that the saline alluvial silty-clayey Serozems received their salts from this source.

Soil Formation from the Lacustrine Marl

The soils of the central part of the valley have been formed from marly-lake deposits (Dan and Alperovitch, 1971). These soils have an AC horizon sequence and are highly calcareous and saline. A gypsic horizon is found as a rule at shallow depths. These soils were defined as saline gypsiferous highly calcareous Serozems. (Typic Torriorthids, sometimes Gypsiorthids or Calciorthids).

The marly terrace was dissected by the Jordan River and its tributaries, due to the gradual drop of the base level of the Dead Sea that started about 12,000 years ago. As a result, badlands were formed in the transitional area between the terrace and the recent floodplain. The soils on these badlands are very shallow and include marly saline desert Lithosols (Typic Torriorthents). In the depression between the badlands, the drainage conditions are very poor and the soils are severely salinized. As a result,

various Solonchaks (Salorthids), both from Lisan marls and alluvial materials, are found in these areas.

Soils of the Jordan Floodplain

The soils in the recent Jordan floodplain are being formed from alluvial material of the Jordan (Dan and Alperovitch, 1971). Here are found various brown Alluvial soils (Torrifluvents or Xerofluvents) in well-drained positions and alluvial Solonchaks (Salorthids) in poorly drained areas. It should be pointed out that the drainage condition becomes gradually poorer toward the Dead Sea, and as a result the Solonchaks cover the whole floodplain in the lowest part of this region. In this area, even sterile alluvial Solonchaks that have been formed by extreme salinization are found.

Soil Formation in the Mildly Arid Parts of the Desert Fringe Areas in the Mountains of Judea and Samaria

Soils of Mountainslopes

The primary stages of soil formation usually resemble those of the more arid mountain areas, especially in the southern parts of this region. Brown Lithosols (Lithic Torriorthents) are formed in pockets among hard rocks, while Rendzinic desert Lithosols (Lithic or Typic Torriorthents) are found on chalks and marls. However, the soils are somewhat more developed and usually also more leached.

The Brown Lithosols in this region are nonsaline and contain usually somewhat more organic material than in the drier regions. In the moister parts of this region these Lithosols merge with Brown Rendzinas (Lithic Xerochrepts or Lithic Haploxerolls) that are darker colored and contain more organic material. In the northern parts of the Judean Desert and especially in the Samarian Desert, the soils on hard rocks are usually fine textured, as the aeolian dust that reached this area contained more clay. As a result, clayey Brown Rendzinas are usually formed in this region.

The leaching in the Samarian Desert is more pronounced because the rate of dust deposition that reaches this area is much less than that in the Negev (Yaalon and Ganor, 1975). As a result these soils are less calcareous, and in favorable spots all the carbonates have been leached and Terra rossa soils (Lithic Xerochrepts) formed (Zeidenberg and Dan, 1979; Litaur, 1980).

On moderate slopes, where the erosion is not severe, the aeolian dust cover gradually becomes thicker and loessial light Brown soils (Typic Haploxeralfs or Calcic Palexeralfs) are formed in the northern Negev and the southern part of the Judean Desert (Dan and Raz, 1970). Transitional soils are found on somewhat steeper slopes. The loessial light Brown soils are especially widespread on northern slopes, where erosion is less severe due to the denser vegetational cover (Yaalon and Dan, 1974).

Soils of Lower Slopes, Valleys, and Depressions

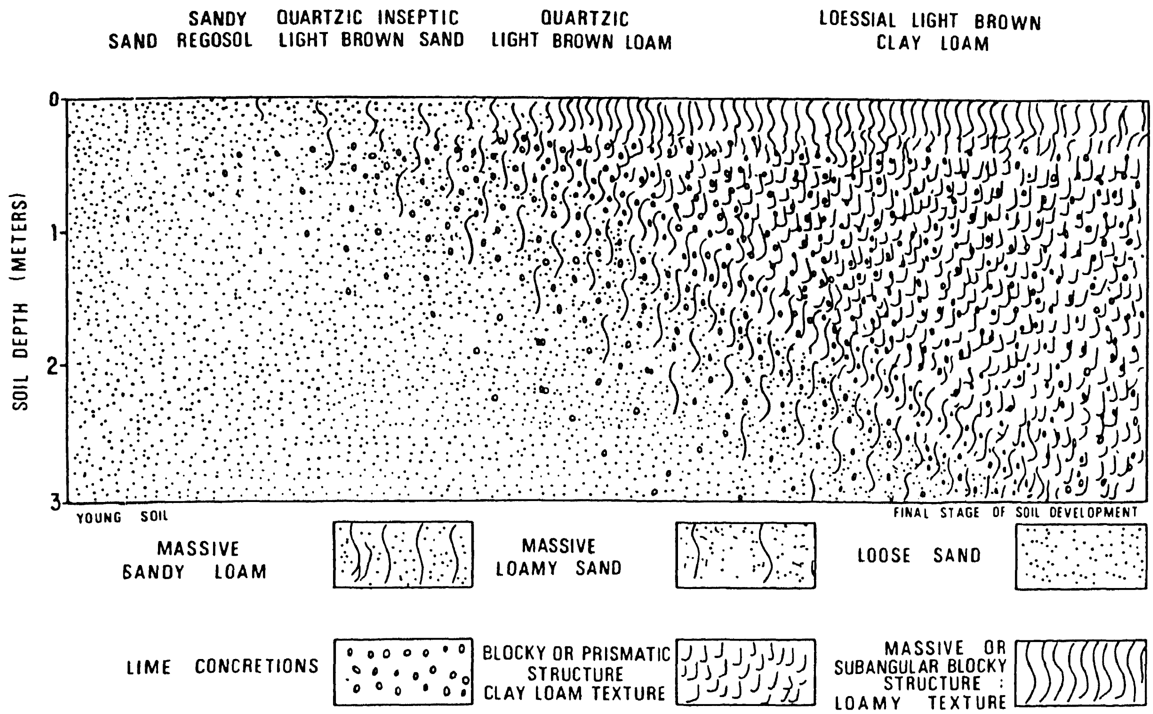
On the lower parts of the slopes and on footslope positions, loessial soil material that is mixed with gravel and stones has been deposited. Soil formation in these places is expressed mainly by formation of a cambic and, later on, even an argillic B and enrichment of lime at a depth of about 50-100 cm. This lime forms a typical calcic horizon. The lime concentration in this layer increases gradually until petrocalcic horizon is formed (see Fig. II.12).

In the northern part of the Judean Desert and in the Samarian Desert, the deep soils that form on moderate slopes, plateaus, and depressions are finer textured, due to the clayey nature of the aeolian dust in this region (Yaalon and Dan, 1974; Dan and Alperovitch, 1975). Natric grumic Serozems (correlated with typic Natrargids) are found in the drier parts of this region, while toward the moister parts these soils merge with Natric Grumusols and finally even with typically calcareous reddish Brown Grumusols (or typic Chromoxererts).

The leaching of these soils is somewhat inhibited due to the fine texture, and as a result some of the salt and large amounts of abasorbed sodium remain in the deeper soil layers. The saline soils are designated as Serozems and, due to the high absorbed sodium, they are defined as Natric Serozems (Committee on Soil Classification in Israel, 1979). The leaching increases gradually toward the moister regions and as a result the salts and farther on also the adsorbed sodium are leached into the deeper soil layers.

The difference between the Grumic soils and the Grumusols stems from differences in the silt:clay ratio of these soils. The proportion of silt in the dust that reaches the soils which are nearer the desert is relatively higher (Yaalon and Ganor, 1975) and, as a result, pedoturbation is inhibited to some extent and natric

Fig. II 13. Soil development from sand in the mildly arid part of the Northern Negev (according to Dan, 1980).



Grumic Serozems were developed (Zeidenberg and Dan, 1978). The clay content increases with increasing distance from the desert, and as a result Grumusols (Typic Chromoxererts) were formed in these areas (Yaalon and Dan, 1974; Dan and Alperovitch, 1975).

Soil Formation in the Mildly Arid Plains of the Northern Negev

The primary topography of this region is governed mainly by ancient dune ridges and gravelly coastal sediments. These sediments were affected by the slow encroachment of aeolian loess.

The youngest soils have been developed mainly from the dune sand; these include sandy Regosols (Quartzipsamments) and quartzic inseptic light Brown sand (also Quartzipsamments merging with Calcixerollic Xerochrepts) which is chracterized already by a Bca horizon (Fig. II.13). With continuous development, the soil and especially the Bca horizon become finer textured and loamy quartzic light Brown soil (Calcic Haploxeralfs) are formed (Dan and Yaalon, 1980).

These soils are characterized already by an argillic Bca horizon. Loess acretion continues until the sandy sediments are completely covered by the loess (Dan and Yaalon, 1971, 1980). Soil formation continues simultaneously, so that a typical cumulic loessial soil has been formed. This soil is characterized by an ochric loamy A horizon and an argillic, clay loamy B horizon. This soil is designated as loessial light Brown clay loam (correlated with calcic Haploxeralfs).

Loess deposition affected this area throughout several hundred thousand years. As a result, a column of about 8-12 meters of clays or silty clays with up to six paleosols, underlies the recent soil (Dan and Yaalon, 1971; Bruins, 1976). Rainfall in this area is high enough to leach most of the soluble salts from the soil profile, but some high ESP values are still found in the deeper soil layers, indicating restricted leaching.

In depressions and floodplains, fluvial redeposition of the loess occurs (Dan et al., 1976; Dan and Yaalon, 1980). Soils are as a rule young, without profile differentiation, and are designated as Loess or alluvial Loess (correlated with typic Xerofluvents). In recent to late Pleistocene terraces, or on footslopes, a cambic loamy Bca horizon has been formed, and the soils are defined as loessial inseptic light Brown loams (correlated with typic or calcixerollic Xerochrepts).

Soil Formation in the Sandy Areas of the Western Negev

The western Negev has been covered recently by aeolian sand that reached the region from the drier area in the south (Dan 1979a,b). These sands covered older soils, mainly quartzic light Brown soils and Loessial light Brown soils (calcic Haploxeralfs). The deposition of this sand has been quite gradual, and therefore the area has been covered by quite a dense vegetation. The ecological conditions in these sandy areas are better than those of the loessial areas farther to the east, due to the absence of runoff and the low waterholding capacity of the sand. Penetration of the rainwater in this area is quite deep, in a normal year reaching a depth of about 1 meter or even more.

Soil development is negligible due to the short time that passed from the time of deposition, and as a result the soils were defined as sandy Regosols (Torripsamments or Quartzipsamments). However, in several places some lime segregation has already been detected, and it seems that sandy quartzic inseptic light Brown soils (Calcixerollic Xerochrepts) are slowly developing in these areas.

Shifting sand dunes are found along a strip several kilometers wide, near the coastline. The sand dunes are usually bare of vegetation, but in depressions between the dunes some vegetation has been found. These depressions are especially noticeable near the seacoast, where their elevation reaches only several meters above sea level and deep-rooted plants can reach the groundwater.

A typical hydromorphic vegetation exists in some of these areas. Bedouins took advantage of this feature and removed sand from smallplots to an elevation of about 1 meter above the water table. On these areas they planted various orchards and vegetables, so that the roots reach the groundwater. The sand removed is piled high on the sides of these plots, thus giving the appearance of sunken gardens.

In conclusion, it should be stressed again that all the sands in this area include only recent sands. In the future these sands will probably form the typical light Brown soils that characterize this climatic zone, as quite dense vegetation usually covers the sand after some time, and the loess deposition may be protected by this vegetation. A pronounced difference thus exists between this zone and the sandy areas in the more arid parts, where the vegetation is too sparse to stop the movement of the sand and where, as a result, soil development will not be marked even after a long period.

Soil Formation in the Mildly Arid Part of the Jordan Valley (Bet She'an Valley and Biqat Kinarot)

As in the lower parts of the Jordan Valley, the soils have been formed mainly either from alluvial material of the surroundidng mountains or from the marly lake deposits (Lisan marl).

Soil Formation from Clayey Alluvial Sediments

The mountains on both sides of the valley are found in the semi-arid and subhumid climatic regions (Rosenan, 1970), and the soils of these regions consist of typical fine-textured clayey Terra rossa, Brown Rendzinas, Protogrumusols, (mainly Lithic Xerochrepts) (Chromoxererts and Grumusols) (Dan and Raz, 1970; Dan et al., 1976). As a result, the alluvial material that reaches the valley consists also mainly of montmorillonite clays, and various Grumusols (Vertisols) have been formed from these sediments.

Brown and Reddish brown Grumusols (Chromoxererts) were found in the well-drained areas. Many of these Brown Grumusols suffer from high ESP values in the deeper soil layers, as rainfall is not sufficient to leach the sodium from the fine-textured soils. These soils are thus defined as Brown and Reddish brown natric Grumusols (Committee on Soil Classification in Israel, 1979). The natric Grumusols are especially widespread in the drier parts of this region, such as in the Bet She'an Valley.

Various Hydromorphic Grumusols (correlated with the Pelloxererts) are also widespread. They are found mainly near the springs and the formerly swampy areas. Large areas of hydromorphic Grumusols are found on the transitional strip between the Grumusols and the soils that were formed from the Lisan marl. The groundwater in this transitional area flowed on the Lisan marl sediments, and as a result the soils suffered from impeded drainage conditions.

Many of these soils contain large amounts of carbonate, as they were mixed to some extent with the underlying highly calcareous sediments; these soils are defined as highly calcareous hydromorphic Grumusols. In some of the soils in this transition zone the hydromorphic features are no longer evident, as they are now well drained. These soils are defined as highly

calcareous marly Grumusols (Committee on Soil Classification in Israel, 1979).

Hydromorphic natric Grumusols are sometimes found in swampy areas. These soils were formed mainly in places where the groundwater contained considerable amounts of soluble salts (Alperovitch and Dan, 1972).

On the upper parts of the alluvial fans, where the parent material contains large amounts of stones and gravel, various Colluvial-alluvial soils (mainly Xerofluvents) are found. In the lower parts they grade with the Brown and Reddish brown Grumusols. On the transitional parts where the amount of gravel and stones is intermediate, gravelly Brown and Reddish brown Grumusols are formed.

Soil Formation from the Lacustrine Lisan Marl

Highly calcareous inseptic Brown soils (included among the calcixerollic Xerochrepts) have developed from the Lisan marl sediments. These soils are characterized by a CA horizon at a depth of about 1/2 to 1 meter. They differ from the saline gypsiferous Serozems of the lower part of the valley by the absence of soluble salts, due to the relatively more intensive leaching (Committee on Soil Classification in Israel, 1979). Hydromorphic highly calcareous inseptic Brown soils were formed in formerly impeded drainage areas that were found among the highly calcareous inseptic Brown soils. Some of these soils contained, in the past, considerable amounts of soluble salts, but most of them have recently been removed, due to the construction of drainage systems that resulted in their being leached by irrigation water.

Pale Rendzinas (Lithic Xerorthents) and Rendzinic desert Lithosols (Lithic and typic Torriorthents) are found on the terrace scarp between the Lisan terrace and the recently formed Jordan floodplain. These soils are very young and do not reveal any diagnostic horizon except for an ochric A horizon, and in the Pale Rendzinas sometimes even the beginning of a mollic epipedon.

Soils of the Jordan Floodplain

The soils of the Jordan floodplain have been formed from recent alluvial material. They are very young and are still characterized by alluvial layering. The texture of these soils grades mostly from loams and silt loams to silty clay loams. These soils are calcareous but usually not saline and thus they were included among the brown Alluvial soils (Typic Xerrofluvents) (Committee on Soil Classification in Israel, 1979).

References

Alperovitch, N. and Dan, J. (1972) Sodium affected soils in the Jordan Valley. Geoderma 8: 37-57.

Amiel, A.J. and Friedman, G.F (1971) Continental Sabkha in the Arava Valley between the Dean Sea and the Red Sea. Bull. Am. Ass. Petrol. Geol. 55: 581-592.

Bruins, H. J. (1976) The Origin, Nature and Strategraphy of Paleosols in the Loessial Deposits of the N.W. Negev (Netivot, Israel). M. Sc. thesis, The Hebrew University of Jerusalem, Jerusalem.

Committee on Soil Classification in Israel. (1979) The Classification of Israel Soils. Spec. Publ. Agric. Res. Orgn, Bet Dagan 137. (Hebrew, with English summary)

Dan, J. (1977) The distribution and origin of Nari and other lime crusts in Israel. Israel J. Earth-Sci. 26: 68-83.

Dan, J. (1979a) The soil pattern of the arid regions of Israel and its effect on present and potential land use. in: Golany, G. {Ed.} Arid Zone Settlement Planning, the Israel Experience. pp. 214-252. Pergamon Press, New York, NY.

Dan, J. (1979b) {The soils of the Negev.} in: Shmueli, A. and Grados, Y. {Eds.} {The Land of the Negev, Man and Desert.} pp. 203-218. Publishing House, Israel Ministry of Defense, Tel Aviv. (in Hebrew)

Dan, J. and Alperovitch, N. (1971) The Soils of the Middle and Lower Jordan Valley. Prelim. Rep. Valcani Inst. Agric. Res., Bet Dagan 694. (Hebrew, with English summary)

Dan, J. and Alperovitch, N. (1975) The origin, evolution and dynamics of deep soils in the Samarian Desert. Israel J. Earth-Sci 24. 57-68.

Dan, J., Gerson, R., Koyumdjisky, H. and Yaalon, D.H. (1981) Ardic soils of Isreal, properties, genesis and management. Spec. Publ. No. 190 Agric. Res. Orgn. Bet Dagan, Isreal.

Dan, J., Moshe, R. and Alperovitch, N. (1974) The soils of Sede Zin. Israel J. Earth-Sci. 22. 211-227.

Dan, J. Nissim, S. (1976) Soils of the Mountainous Parts of the Judean and Samarian Deserts. Spec. Publ. Agric. Res. Orgn, Bet Dagan. 60. (Heberew, with English summary)

Dan, J. and Raz, Z. (1970) The Soil Association Map of Israel (scale 1:250,000). Volcani Institute of Agricultural Research, Bet Dagan; and Dept. of Soil Conservation, Tel Aviv. (Hebrew, with English summary)

Dan, J. and Yaalon, D.H. (1971) On the origin and nature of the paleopedological formations in the coastal desert fringe areas of Israel. in: Yaalon, D.H. {Ed.} Paleopedology - Origin, Nature and Dating of Paelosols. pp. 245-260. Israel Universities Press, Jerusalem.

Dan, J. and Yaalon, D.H. (1980) The evolution of soil and landscape in the northern Negev. Studies in the Geography of Israel 11: 31-56.

Dan, J., Yaalon, D.H., Koyumdjisky, H. and Raz, Z. (1976) The Soils of Israel (with map 1:500,000). Pamph. Agric. Res. Orgn, Det Dagan 159.

Dan, J., Yaalon, D.H., Moshe, R. and Nissim, S. (1982) Evolution of Reg soils in Isreal and Sinai. Geoderma 28: 173-202.

Danin, A., Orshan, G. and Zohary, M. (1975) The vegetation of the northern Negev and the Judean Desert of Israel. Israel J. Bot. 24: 118-172.

Evenari, M., Shanan, L. and Tadmor, N. (1971) The Negev: The Challenge of a Desert. Harvard University Press, Cambridge, MA.

Horowitz, A. (1979) The Quaternary of Israel. Academic Press, New York, NY.

Kovda, V.A. (1973) Soils in relation to salinity, irrigation and drainage. in: Irrigation, Drainage and Salinity - An International Source Book. FAO/UNESCO. pp. 55-79. Hutchinson, London.

Litaur, M. (1980) The soils in a variegated region of northeasatern Samaria: their character, means of formation and influence on plant associations. M.Sc. thesis, Tel Aviv University, Ramat Aviv, Israel. (Hebrew, with English summary)

Picard, L. (1943) Structure and Evolution of Palestine. Bull. Dept. of Geology, The Hebrew University of Jerusalem, Jerusalem.

Picard, L. (1970) Geological Map of Israel, 1:500,000. Atlas of Israel, III/1. Survey of Israel, Ministry of Labour, Jerusalem; and Elsevier Co., Amsterdam.

Ron, Z. (1967) {The Geomorphology of the Elat Region, including a 1:100,000 Map.} Survey of Elat Region, Israel. pp. 83-108. The Elat Regional Council. (in Hebrew)

Rosenan, N. (1970) Rainfall Map of Israel. Atlas of Israel, IV/.2. Survey of Israel, Ministry of Labour, Jerusalem, and Elsevier Co., Amsterdam.

Soil Survey Staff (1975) Soil Taxonomy Handbk. U.S. Dept. Agric. 436.

Tsoar, H. (1978) Dynamics of Longitudinal Dunes. Ph.D. thesis, The Hebrew Univ. of Jerusalem, Jerusalem. (Hebrew, with English summary)

Waisel, Y., Pollak, G. and Cohen, Y. (1978) {The Ecology of Vegetation of Israel.} Div. of Ecology, Tel Aviv University, Ramat Aviv, Israel. (in Hebrew)

Yaalon, D.H. (1963) On the origin and accumulationof salts in groundwater and in soils of Israel. Bull. Res. Coun. Israel 11G: 105-131.

Yaalon, D.H. (1978) Geoderma - Continental Sedimentation, Calcrete, Desert Loess and Paelosols, Sand Dunes and Eolianites. Guidebook 10th Int. Congr. Sedimentology, Jerusalem. part II, pp. 195-238.

Yaalon, D.H. and Dan, J. (1974) Accumulation and distribution of Loess-derived deposits in the semi-desert and desert fringe areas of Israel. Z. Geomorph. (Suppl.) 20: 91-104.

Yaalon, D.H. and Ganor, E. (1975) Rates of aeolian dust accretion in the Mediterranean and desert fringe environments of Israel. Int. Congr. Sedimentology, Nice. 2: 169-174.

Zeidenberg, R. and Dan, J. (1981) The Influence of Lithology, Relief and Exposure on the Soil and Vegetation of the Arid Region of Eastern Samaria. Israel J. Bot. 30: 13-31.

Zohary, M. (1955) {Geobotany.} Sifriat Hapoalim, Merhavya. (in Hebrew)

Physical Properties Affecting the Productivity and Management of Clay Soils in Saskatchewan

E. de Jong* and J.A. Elliott[1]

Abstract

In Saskatchewan, clay and heavy clay textured soils historically are rated as having a higher productivity than light- and medium-textured soils. Data from field experiments are used to illustrate how differences in soil water, temperature, aeration, and mechanical properties of the root zone could lead to differences in productivity for soils of different texture. The higher productivity of clay soils probably is due largely to their higher water holding capacity (an important factor in the prevalent crop-fallow rotation), possibly higher infiltration of snowmelt water, and higher hydraulic conductivity at the permanent wilting point. Higher denitrification losses could explain why yields on clay soils peak at lower total crop water use than yields on medium-textured soils. Problems with wind erosion and lack of spring workdays may be more severe on clay soils than on medium-textured soils.

Introduction

This paper compares the physical properties of clays and medium-textured soils, with particular reference to those properties that affect on the agronomic potential of these soils. Field data or data obtained on undisturbed core samples are used as much as possible. Unfortunately, the records often cover only a few years and frequently different soils were involved.

Four soil physical factors are recognized as controlling the productivity and management of soils: the soil water, the soil aeration and the soil temperature regimes, and the mechanical properties of the root zone. These four factors vary over space and time and to a degree depend on each other. In particular, variations in water content affect the aeration, temperature, and mechanical strength of a soil.

Productivity

Historical yield records indicate that the heavier-textured glacial lake deposits represent some of the best agricultural land in Saskatchewan (see Table 17 in Mitchell et al., 1944). The high productivity rating of these soils reflects their high fertility and drought resistance and the fact that their smooth and undulating topography and absence of stones make cultivation easy. Drainage is generally adequate, although temporary flooding of flat and depressional areas occasionally interferes with farm operations and drowns crops.

Spring wheat is the most commonly grown arable crop in Saskatchewan, and the productivity rating in Mitchell et al. (1944) is based mainly on wheat grown on fields that were fallowed the previous year. Under these conditions, the higher water storage capacity of the heavier textured soils is a distinct advantage. When yield is plotted against crop water use (Fig. 1), the difference in maximum yields obtained on heavy and medium-textured soils is small and probably not significant. Figure 1

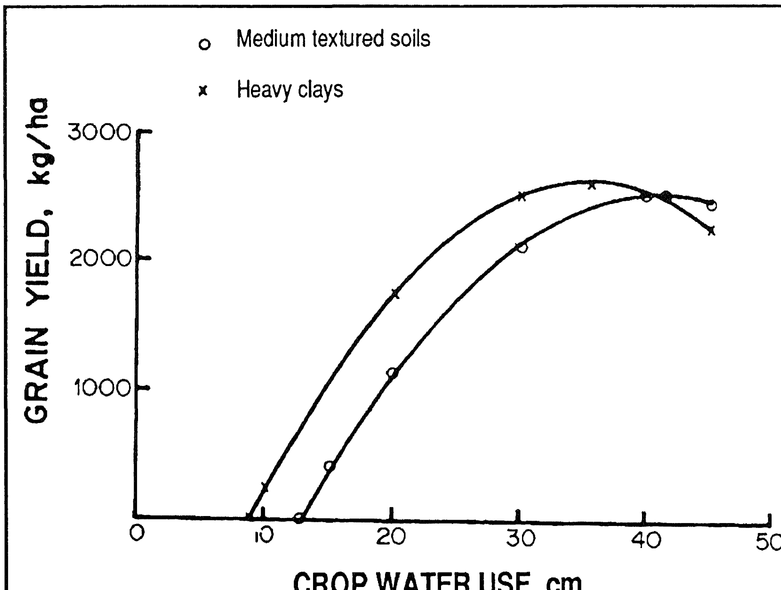


Fig. 1 Relationships between spring wheat yields and water use for Brown and Dark Brown soils (de Jong and Halstead, 1987).

[1]E. de Jong and J.A. Elliott, Saskatchewan Institute of Pedology, University of Saskatchewan, Saskatoon, Saskatchewan S7N OWO, Canada.
*Corresponding author.

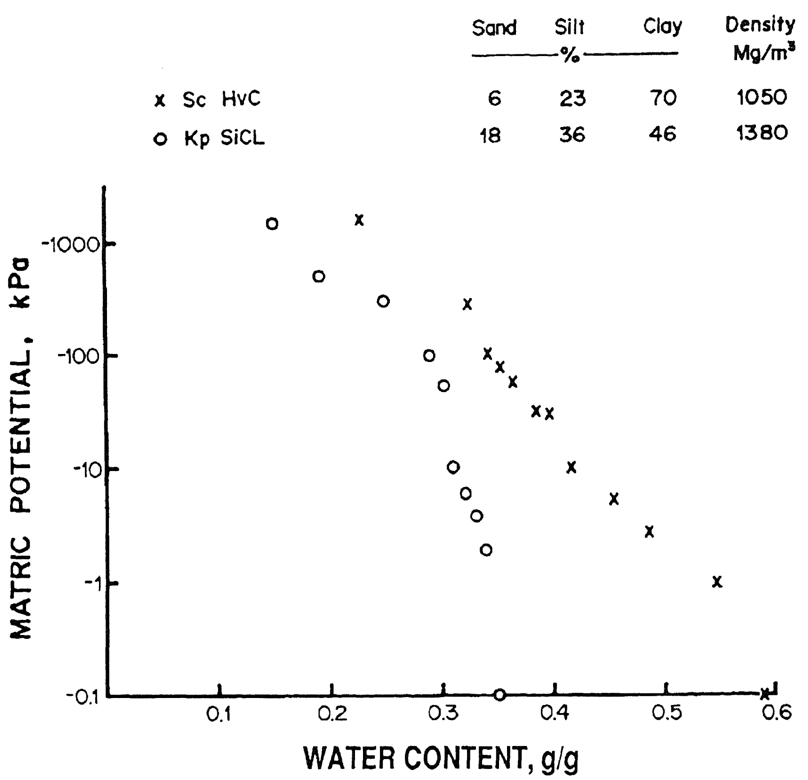


Fig. 2 Water retention of the subsoils of a Sceptre heavy clay (de Jong and Stewart, 1973) and Keppel silty clay loam (Patterson, 1985).

does illustrate that the heavier soils yield better at low water use than the medium-textured soils, while at high water use the reverse is true. Thus, under semiarid conditions and when practicing continuous cropping, the crop uses water more efficiently on heavier than lighter textured soils. In the following sections, some of the factors responsible for this difference are explored.

Soil Water Regimes

The amount of water held at a given matric potential generally increases as clay content increases (Fig. 2). When the water content is expressed on a volume basis, the differences are less pronounced as the bulk density generally increases with decreasing clay content. Whether on a volume or a weight basis, available water holding capacity, the difference between field capacity (assumed to correspond to -33 kPa matric potential) and permanent wilting point (-1500 kPa matric potential), generally increases with increasing clay content. The changes in soil volume that accompany changes in soil water content result in shrinkage cracks that can have a significant effect on transport processes in clay soils.

The saturated hydraulic conductivity of clay soils is generally lower than that of medium- and light-textured soils. For example, the saturated hydraulic conductivity of the Sceptre soil in Figure 2 is about 30-40 cm/day, while the value for the Keppel soil is about twice that. However, as the soil water content decreases, the unsaturated hydraulic conductivity of a clay soil decreases less rapidly than that of a lighter soil (Fig. 3; see also Gardner, 1960; Yang and de Jong, 1972). At -33 kPa matric potential, both soils have a hydraulic conductivity of about 10^{-2} cm/day, but at -1500 kPa the heavy clay has a hydraulic conductivity of 10^{-4} cm/day compared to around 10^{-5} cm/day for the silty clay loam. The higher hydraulic conductivity of the clays at -1500 kPa may explain their higher productivity than lighter soils when water is limiting (Fig. 1) and lowers the matric potential at which permanent wilting occurs (Yang and de Jong, 1972).

Hydraulic conductivity is highly variable (Warrick and Nielsen, 1980). Data from Kirkland (1986) shows that the hydraulic conductivity of a shrinking clay was more variable than that of a non-shrinking medium textured soil (Fig. 4). The hydraulic conductivities were measured on undisturbed soil cores collected at various times

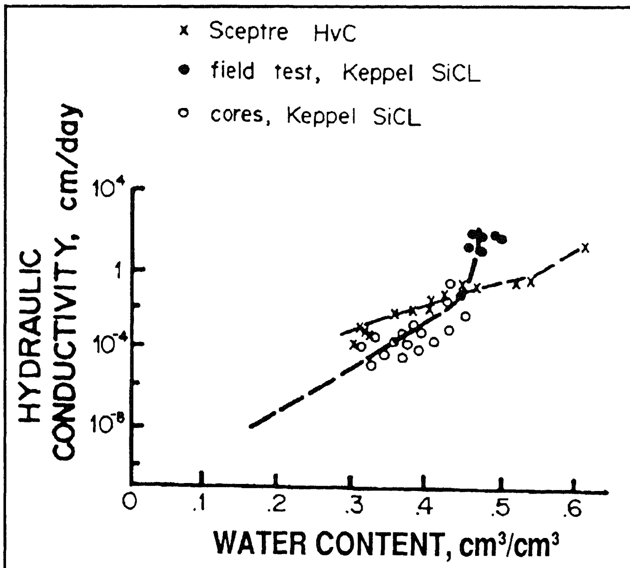


Fig. 3 Dependence of hydraulic conductivity on soil water content for Sceptre heavy clay (de Jong and Stewart, 1973) and Keppel silty clay loam (Patterson, 1985).

during three growing seasons (each value is the geometric mean of two or more cores). The higher variability in the clay is probably due to the presence or absence of shrinkage cracks which, even though closed when the soil is saturated, represent preferential flowpaths (Ritchie et al., 1972; Bouma, 1981). At -5 kPa matric potential, the variability had significantly decreased, as cracks or pores larger than 60 µm would no longer contribute to water flow and, hence, the major source of the variability was reduced.

Cracking in clay soils increases infiltration rates (Allan and Braud, 1966; Blake et al., 1973). Large amounts of water more or less instantaneously disappear into the cracks when water is ponded on the surface (Fig. 5). Johnson (1962) has suggested that vegetation-induced, controlled soil cracking might be an economically feasible way to increase infiltration on dry-farmed wheatlands. In Saskatchewan these cracks would be important only for infiltration during the rare, intense summer storms which might cause surface ponding. At low and moderate rainfall rates the soil would gradually wet up, and rain falling directly in the cracks probably would be absorbed by the walls.

On the Canadian Prairies, infiltration of water from the melting snowpack is often a major soil water recharge event, and cracks do contribute significantly to the entry of snowmelt water. In non-cracked soils, snowmelt infiltration depends on the amount of snow water and the frozen water content of the surface soil (Gray et al., 1984). Snowmelt infiltration in dry, cracked clays usually is limited only by the amount of water in the snowpack. Gray and his co-workers are studying the possibility of increasing snowmelt infiltration by creating artificial cracks. On heavy clay soils, variation in the spacing and orientation of wheat rows with respect to the slope might naturally give the desired effect.

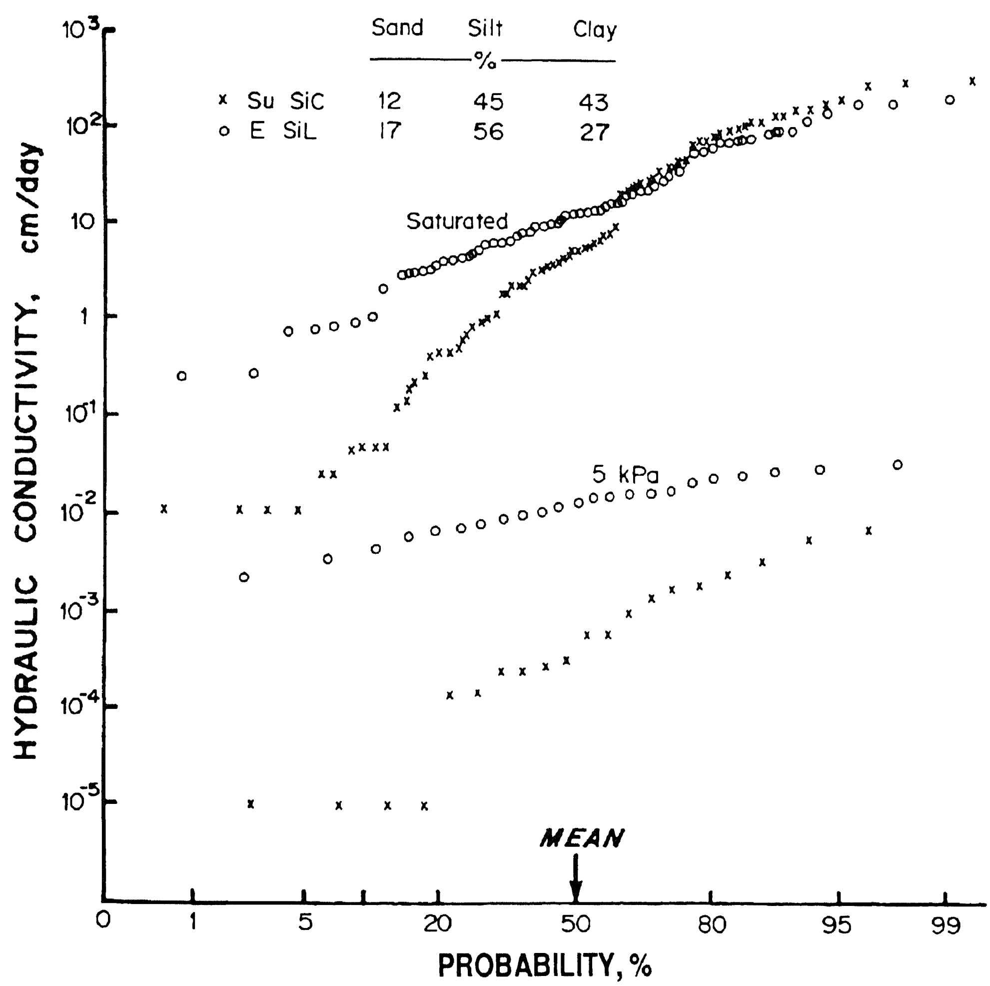


Fig. 4 Hydraulic conductivity of a Sutherland silty clay and an Elstow silty loam (Kirkland, 1986).

Cracks obviously increase the possibility of evaporation losses from the body of the soil. For example, Johnston and Hill (1945) found decreased water contents in the soil adjacent to the cracks. Selim and Kirkham (1970) estimated that cracks increased evaporation by up to 30% in a laboratory experiment; these increases are probably larger than those which occur in the field as the soil was saturated initially. The work of Selim and Kirkham suggests that several small cracks probably lead to greater water losses than a single, wide crack.

The soil water regime is a function of

Table 1. Maximum and minimum soil water contents under different vegetation on Sceptre heavy clay (from de Jong and MacDonald, 1975).

Vegetation	Total water to 135 cm depth Maximum	Minimum
	------ cm ------	
Native grassland	55	38
Wheat on fallow	58	41

Table 2. Estimated range of available water (50% probability) at seeding and heading for two soil types (adapted from de Jong and Bootsma, 1988).

Area	Northeast Saskatchewan		Southwest Saskatchewan	
Rotation	Wheat/Fallow	Cont.Wheat	Wheat/Fallow	Cont.Wheat
	mm available water			
At seeding				
Heavy class[1]	240-210	210-140	140-120	90-80
Medium texture[2]	140-125	135-105	95-80	65-55
At heading				
Heavy clays[1]	150-125	115-95	80-55	55-35
Medium texture[2]	65-55	55-45	35-25	25-15

[1]250 mm water holding capacity
[2]150 mm water holding capacity

vegetation, topography, soil type, and climate. The effect of vegetation is illustrated in Table 1. Continuous wheat would be expected to have water contents between those of the native grassland and the wheat grown on summerfallow (Table 1). Kirkland (1986) found little difference in the water extraction pattern of continuous wheat grown on Sutherland silty clay and Elstow silty loam, under two tillage systems (Fig. 6). Note that overwinter-recharge was similar on both soils in 1983-84 when the soils were wet the previous fall. After the dry summer of 1984, the silty clay gained considerably more water over the 1984-85 winter than the silty loam, presumably due to the role of cracks in snowmelt infiltration. Much of the additional recharge in the Sutherland silty clay was found below 50 cm depth.

Unfortunately there are no continuous long-term records for the soil water regime under different soil/crop combinations. De Jong and Bootsma (1988) have used the Versatile Soil Moisture Budget (VSMB) to predict variability in soil water on the Canadian Prairies. The VSMB provides as good a prediction of growing season changes in soil moisture under Saskatchewan conditions as more sophisticated models (cf. de Jong and MacDonald [1975] versus de Jong and Hayhoe [1984]; de Jong, 1988), but problems tend to arise in the predictions of overwinter soil water gains (de Jong and MacDonald, 1975). Despite this, de Jong and Bootsma (1988) ran the model for 60 years using existing climatic records. Table 2 shows some of their predictions for continuous wheat and wheat on fallow on a heavy and a medium-textured soil. At maturity both the wheat on fallow and the continuous wheat in southwest Saskatchewan would have depleted most, if not all, of the available water. Some available water probably would still be present in both rotations in northeast Saskatchewan, where the climate is less arid. Fallow fields would have water contents similar to fields of continuous wheat in spring and would stay at that water content until about September. From September to the next spring, all fields would recharge to the seeding water contents represented by wheat on fallow and continuous wheat in Table 2.

Dasog (1986) used the predictions of de Jong and

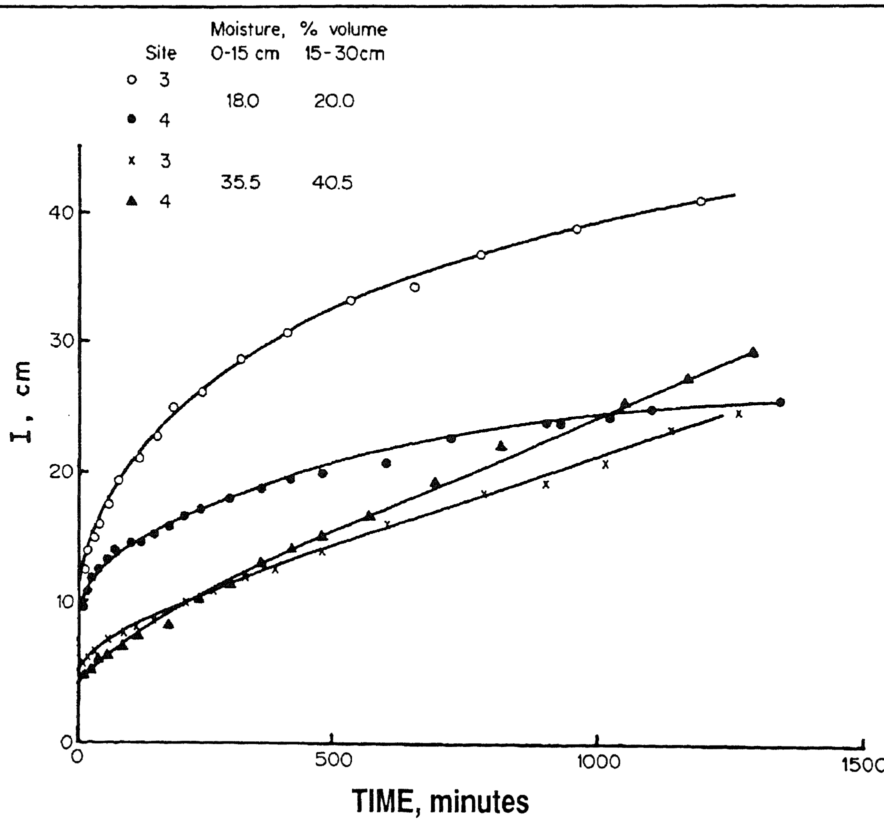


Fig. 5 Cumulative infiltration (I) in Sceptre heavy clay as a function of time at two moisture contents.

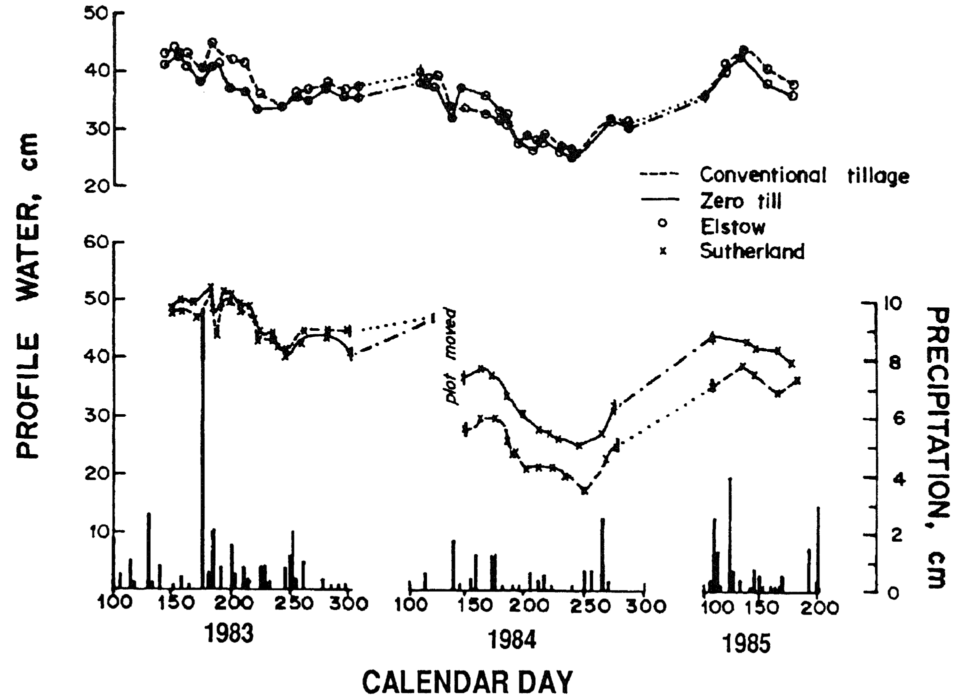


Fig. 6 Temporal changes in profile water under zero and conventional tillage on Sutherland silty clay and Elstow silty loam.

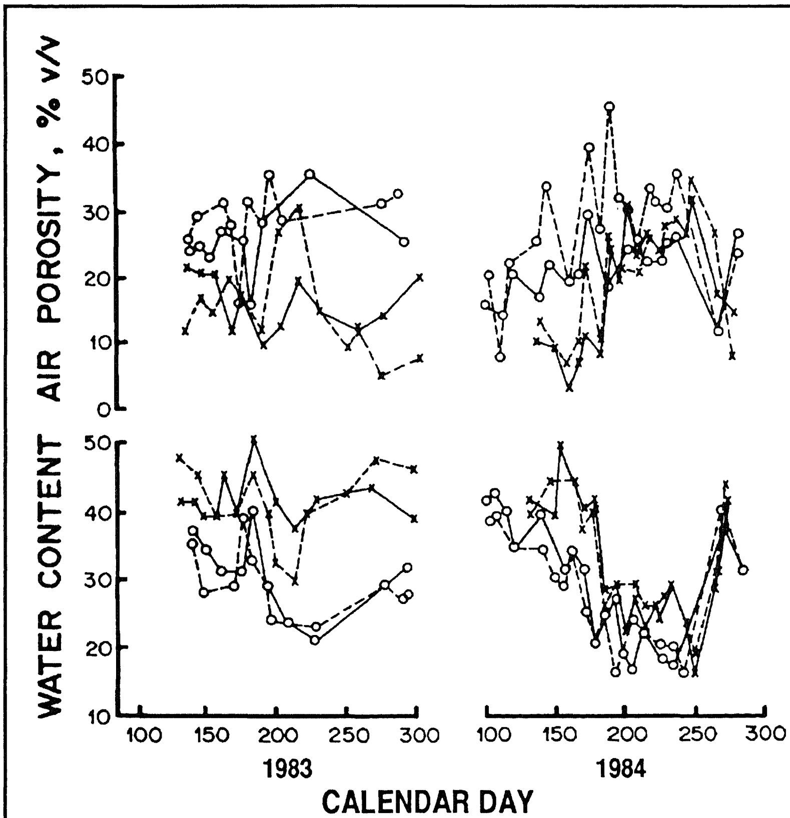


Fig. 7 Surface soil air-filled porosity and volumetric water content for Sutherland silty clay and Elstow silty loam (see Fig. 6 for legend).

Bootsma (1988) to estimate available water at various depths in clay soils and to predict crack duration. The results are summarized in Table 3. Since the model does not take into account the enhanced snowmelt recharge and evaporation associated with cracks, it is doubtful whether the data in Table 3 is particularly reliable. Dasog (1986) did not try to evaluate the cracking outside the growing season, since no available water predictions were tabulated between November 1 and April 1. Dasog's calculations do indicate that cracking is more severe in the subarid southwest than the subhumid northeast of Saskatchewan, and they point out the increased frequency of cracking at depth. The latter is borne out by de Jong and MacDonald (1975), who found that from 1968 to 1972 water contents below 75 cm were virtually steady (at -1500 kPa) under native grassland in southwest Saskatchewan. Although the model may give some indication of the extent of cracking, it does not and cannot predict the number of cracks (i.e., crack spacing).

Soil Aeration

Kirkland (1986) showed no significant differences in soil aeration between zero- and conventional tillage and only a small effect of texture. Surface water contents were generally higher on the Sutherland silty clay than the Elstow silty loam (Fig. 7); therefore air-filled porosity showed the opposite trend. For much of 1983, air-filled porosity was below 20% in the Sutherland soil because of the abundant rains (Fig. 6). Air-filled porosities of less than 10% limit root growth (Wesseling, 1974), although the critical value may be nearly twice as high in Vertisols (Hodgson and MacLeod, 1989). Despite the potential for restricted aeration, in 1983 the aver-

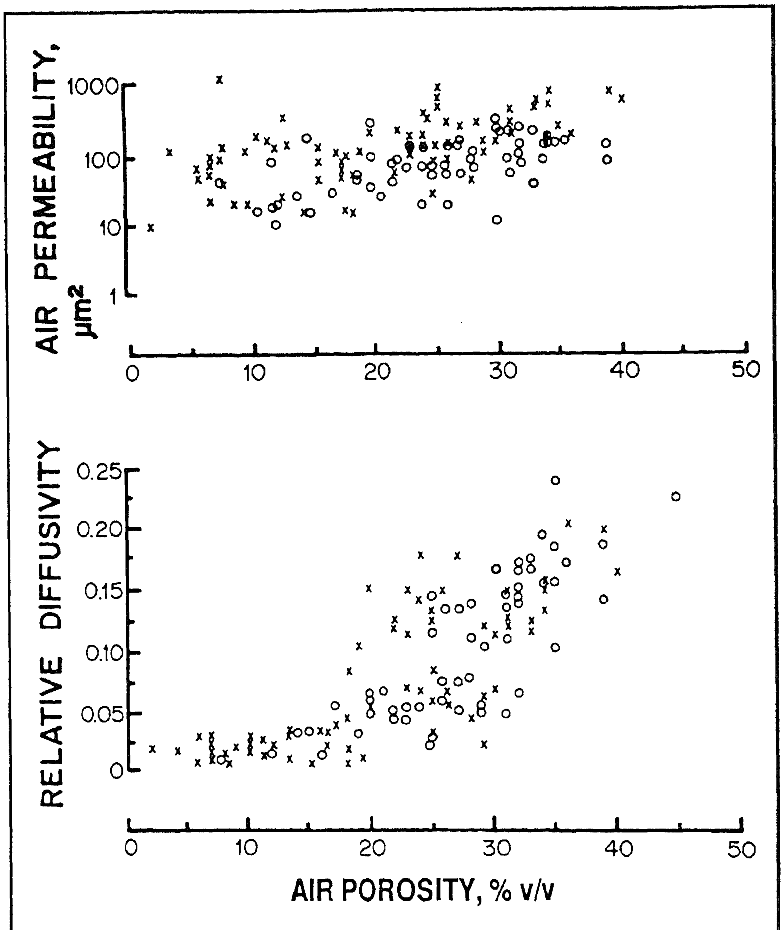


Fig. 8 Air permeability and relative diffusivity as a function of air-filled porosity for Sutherland silty clay (x) and Elstow silty loam (o).

age soil CO_2 concentrations did not exceed 2.5% in the Sutherland soil and 1.5% in the Elstow. CO_2 levels were much lower in the dry 1984 growing season.

To compare pore-efficiency for gaseous transport, air permeability and relative diffusivity were measured several times in the 1982, 1983, and 1984 growing seasons on 7.5 cm diam. by 7.5 cm high soil cores. The relationship between relative diffusivity (the ratio of gaseous diffusion in the soil to diffusion in air) and air-filled porosity was similar for both soils, but air permeability appeared to be slightly higher on the Sutherland silty clay than the Elstow silty loam (Fig. 8). This would indicate that the Sutherland soil has a better developed system of continuous, large, more-or-less vertical pores than the Elstow silty loam. Presumably these large, more or less vertical pores are the shrinkage cracks in the Sutherland soil, even though an effort was made to avoid cracks when the samples were collected. Thus, the effect of cracks on air permeability is probably under-estimated.

Although the composition of the air in the large pores was similar in the silty clay and silty loam, this is not necessarily true for aeration inside the aggregates. In non-shrinking soil, air enters the aggregates as the moisture content drops. Aggregates of shrinking soils exhibit normal shrinkage (i.e., any decrease in water content is accompanied by an equal decrease in volume), as Dasog's data for two heavy clays illustrate (Fig. 9).

Using data from undisturbed cores, de Jong and Stewart (1973) calculated that at saturation the Sceptre subsoil had a bulk density of 1250 Mg/m^3 and a water content of 0.44 g/g. The cores showed structural shrinkage (volume loss less than the volume of water lost) between 0.44 and 0.42 g/g water con-

Table 3. Estimated crack duration (days) from May 1 to Sept. 1 in clay soils with 200 mm available water holding capacity (Dasog, 1986).[1]

Depth (cm)	NortheasSaskatchewan	SouthwesSaskatchewan
	------- days with less than 40% available water -------	
0-6	80	80
6-15	79	86
15-30	71	78
30-60	65	83
60-90	82	105
90-120	90	115

[1] Assuming 200 mm storage and continuously cropped to wheat

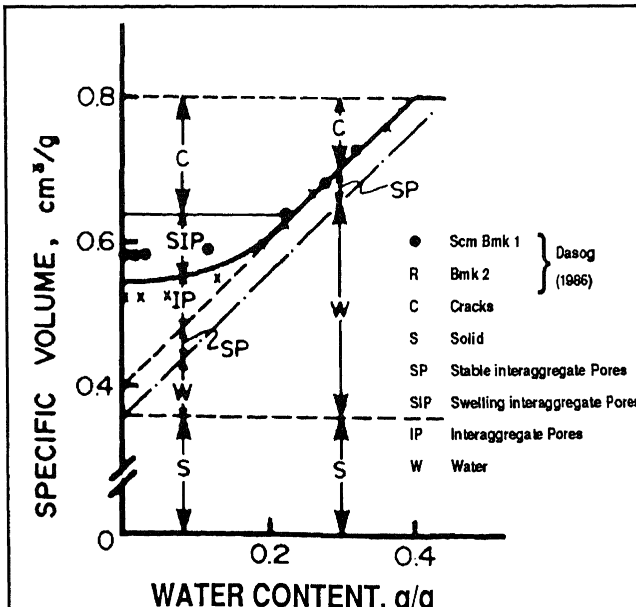


Fig. 9 Pore space in a Sceptre heavy clay.

tent and normal shrinkage from 0.42 to 0.24 g/g water content. During structural shrinkage, air will enter some stable pores in and between aggregates, and air-filled pores (or cracks) between aggregates should increase in volume. The air-filled pores within the aggregates should not increase until residual shrinkage starts (at about 0.22 g/g in Fig. 9). The -1500 kPa water content for disturbed Sceptre soil is about 0.25 g/g, which would suggest that the aggregates remain nearly water-saturated over the range of plant available water.

Because of the difference in degree of water-saturation, anaerobic microsites are more likely to occur in heavier than lighter textured soils, even when the composition of the air in the interaggregate pores is similar. Denitrification measurements support this; e.g., Colaco (1979) measured five times as much denitrification on a fallowed Sutherland clay than on a medium-textured soil similar to the Elstow. The increased denitrification could perhaps explain the lower yields at high water use on heavy clays than on medium-textured soils (Fig. 1).

Soil Temperature

At the same water content (g/g or cm^3/cm^3), lighter textured soils tend to have a higher thermal conductivity and thermal diffusivity than clays (van Wijk and de Vries, 1963). Hence one would expect that in the same climatic setting, and other factors being similar, temperature fluctuations would be slightly larger in the sandier soil, but the means would be quite similar for soils of different texture (cf. Geiger 1959, Table 24). Since heavier textured soils are often wetter than lighter soils, they usually warm up slower in the spring and cool down slower in the fall than lighter soils (Shulgin 1957, Table 20).

Figure 10 compares 10-day running means of soil temperatures in the Sutherland and Elstow soils studied by Kirkland (1986). Tillage treatment only affected soil temperatures in 1984 when the conventional till plot on the Elstow soil was consistently warmer at 0 and 20 cm than its zero-till counterpart. This was probably due largely to less shading by vegetation as the conventional till plot yielded 480 kg grain/ha versus 1135 kg grain/ha for the zero-till plot. The temperature patterns in 1983 follow the expected trend, with the silty clay soil being somewhat colder than the silty loam before mid-July and the reverse being true in the fall. During the very dry 1984 growing season, the soil temperatures at 20 and 50 cm depths in the silty clay were persistently a few degrees lower than in the silty loam. Since moisture conditions were quite similar for both sites (Fig. 6), the lower subsoil temperatures could be due to evaporation from the walls of cracks (Selim and Kirkham, 1970).

Ripley (1973) observed very rapid warming (from -2 to +1°C) of a Sceptre heavy clay soil under native vegetation in spring and suggested that infiltration of the snowmelt water down the cracks was the likely cause. No such trend was observed for the Sutherland silty clay in the spring of 1985, even though there was considerable recharge (Fig. 6), but the trend may not be visible using 10-day means. Also, the major difference in recharge between the two soils was below 50 cm depth, where no temperature data were collected.

From the limited data it appears that the thermal regimes of mineral soils are not strongly affected by texture. It is therefore unlikely that different temperature regimes would account for the differences in soil productivity between heavy and light textured soils.

Mechanical Properties of the Root Zone

Attempts to define the mechanical properties of the root zone that lead to high soil productivity generally have been unsuccessful. Several authors have reported critical bulk densities above which root growth is limited in various soil textures (e.g., Pierce et al., 1983), but few of Saskatchewan's soils exceed the critical limits when sampled at -33 kPa matric potential (Ayres et al., 1973). At -1500 kPa matric potential, clod densities of shrinking soils may approach or exceed the critical values (Dasog, 1986). Others have concentrated on establishing critical soil strengths for root growth (e.g., Taylor et al., 1966), which decreases as clay content increases (Gerard et al., 1982). Ayres et al. (1973) measured the strength of several Saskatchewan soils at -33 kPa and found little difference between heavy and light-textured soils; none of the measured values were close to the critical soil strengths reported by Taylor et al. (1966).

Although attempts to relate aggregate size distribution to crop production have been largely fruitless, aggregate size distribution is important in terms of soil erodibility. Water erosion risk maps for Canada (Coote et al., 1982)

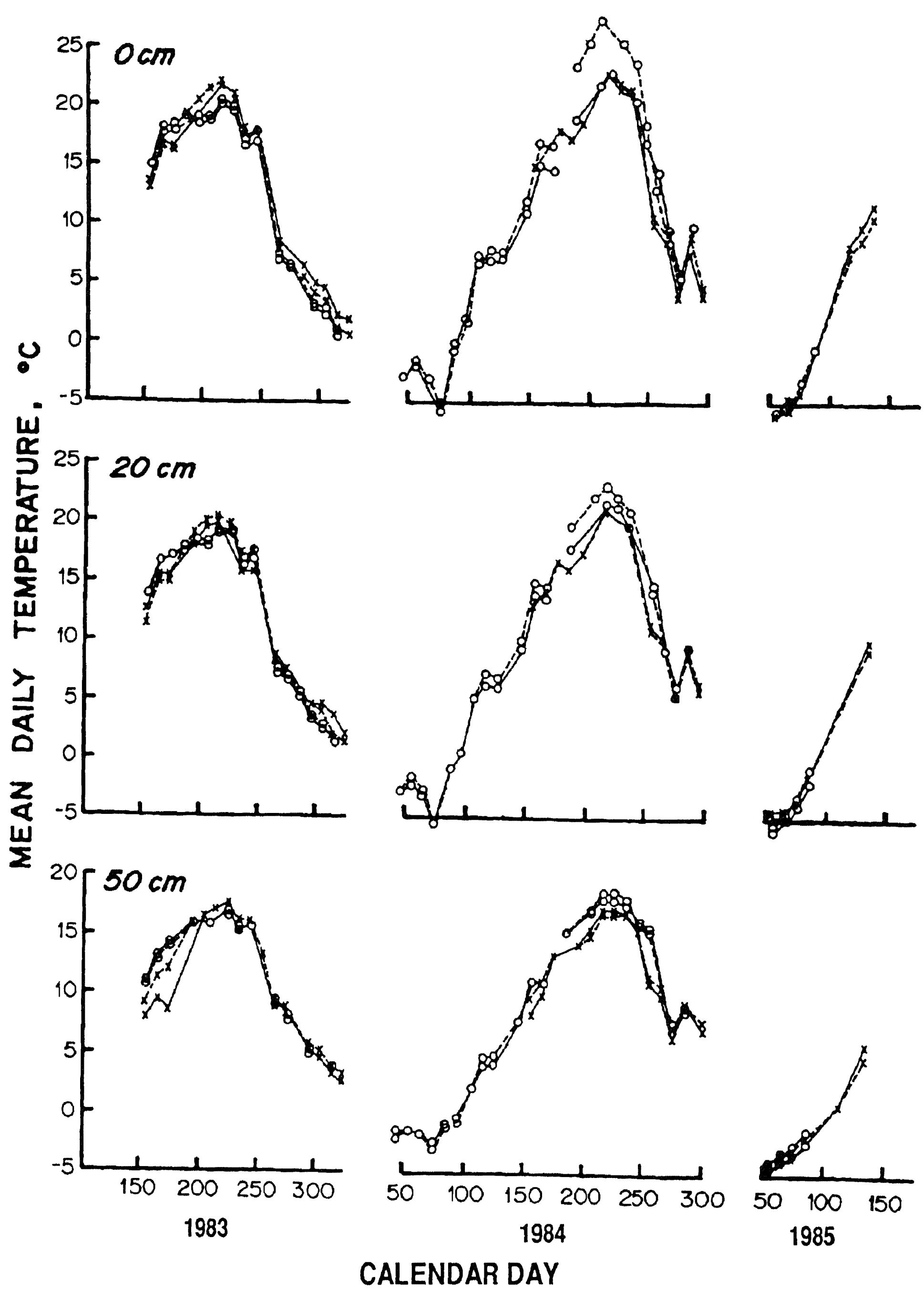


Fig. 10 Mean daily temperature (10 day averages) for Sutherland silty clay and Elstow silty loam (see Fig. 6 for legend).

Although texture may be one indicator of soil erodibility, aggregate size distribution is the critical factor, as both wind- and water-erodibility increase as the percentage of aggregates less than 0.5 to 1 mm increases. Aggregate size distribution changes continually in response to weather and management, and it is impossible to make any firm statements on differences among different soil textures. Figure 11 shows the variation in the erodible fraction (aggregates <0.83 mm diameter as a percentage of the total soil weight) of five different texture groupings. There is a distinct difference in the behavior of the heavier and light-textured soils over the winter period, with the former showing an increase in erodibility due to frost action. The effect of frost on aggregate stability is strongly affected by water content at freezing, initial aggregate size, and how the soils were dried after thawing (Hinman and Bisal, 1968). In winters with little or no snow cover, freeze-drying leads to a very finely granulated structure on clay soils such as the Sceptre and Regina associations and makes them highly erodible.

show an increase in water erosion risk as the clay content of soils increases, even though Evans (1980) indicated that medium-textured soils were most at risk. In general wind erodibility decreases as silt and clay content increase (Wilson and Cooke, 1980; Coote et al., 1982); however, on the Canadian Prairies medium textured soils are considered most resistant to wind erosion (Johnson, 1983; Chepil, 1953).

Going from southwest to northeast Saskatchewan, the length of the frost-free period decreases from about 120 to 100 days, and workability of the soil in the spring becomes critical. High soil water contents are more likely to limit workability on heavy- than light- textured soils. The plastic limit often is used as the maximum

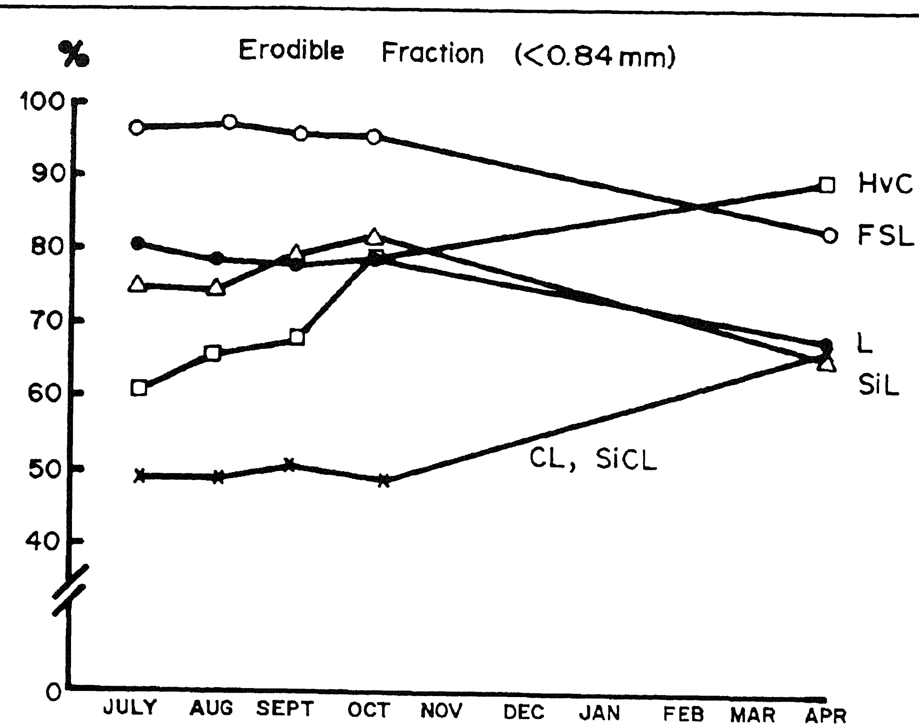


Fig. 11 Erodible fraction (% <0.84 mm) for five soil textural groups from July to April (Hilliard et al., 1988).

water content at which soils should be worked, and its relationship to field capacity (-33 kPa water content) depends on texture. Preliminary analysis of Atterberg limit data for Saskatchewan topsoils suggest that the plastic limit is related to the -33 kPa water content (of disturbed samples) by

$$PL = 2.6 + 0.42 \times H_2O_{-33\ kPa} \qquad R^2 = 0.14$$

Although the equation has a very low coefficient of determination, it suggests that more time is required after the spring thaw for clays to become workable than for lighter textured soils. The importance of texture in determining spring field workdays is illustrated in Table 4. Especially in the subhumid area, the number of working days is substantially less on heavy clays than on lighter soils.

Conclusions

Past experience has shown that clay soils in Saskatchewan are more productive than medium- and light- textured soils, especially under dry conditions and crop-fallow rotations. The higher productivity of the clays is attributed to their higher water holding capacity, which results in increased available water in the root zone of fallow-seeded crops. The cracks in dry clays increase snowmelt infiltration, thus causing higher levels of available water than in lighter soils, in years when abundant snow follows a dry autumn. At -1500 kPa matric potential, the hydraulic conductivities of the clays are higher than those of coarser soils, which may account for the higher water use efficiency of crops grown on clay when water is limiting. Hydraulic conductivity is spatially more variable in clays than in medium-textured soils, probably due to the contribution of cracks to water movement.

The composition of the soil air in clay soils is similar to that in coarser soils, but denitrification data indicate that anaerobic microsites are more prevalent in clays. The higher loss of N by denitrification could explain why maximum yields occur at about 35 cm water use on clays and 40 cm water use on medium-textured soils. Soil temperature regimes appear to be similar for different textures, although infiltration of snowmelt water in the cracks might cause a more rapid thawing in the subsoils of clays than in non-shrinking soils in the spring.

The number of spring working days is less on clays than on medium- and light-textured soils and this can become critical in the subhumid area of Saskatchewan where the growing season is short. Although clays are generally considered to be less erodible than lighter soils, freeze-thaw cycles under snow-free conditions break down the aggregates of clays and make them as wind-erodible as fine sandy loams.

Table 4. Expected (50% probability) spring work days (Dyer et al., 1978).

Location	Soil	April 1 - May 5	April 1 - June 2
		work days	
Regina	Heavy	17.7	35.2
	Light	21.9	46.5
Melfort	Heavy	9.2	33.6
	Light	18.3	42.5

Literature Cited

Allen, J. B. and H. J. Braud, Jr. 1966. Effect of cracks and initial moisture content on the infiltration rate of Sharkey clay. Bull. No. 613, Louisiana State Univ. and Agric. and Mech. College, Agric. Exp. Station.

Ayres, K. W., R. G. Button and E. de Jong. 1973. Soil morphology and physical properties. II. Mechanical impedance and moisture retention and movement. Can. J. Soil Sci. 53:9-19.

Blake, G., E. Schlichting and U. Zimmermann. 1973. Water recharge in a soil with shrinkage cracks. Soil Sci. Soc. Am. Proc. 37:669-672.

Bouma, J. 1981. Soil morphology and preferential flow along macropores. Agric. Water Manage. 3:235-250.

Chepil, W. S. 1953. Factors that affect clod structure and erodibility of soil by wind. I. Soil texture. Soil Sci. 75:473-483.

Coote, D. R., J. Dumanski and J. F. Ramsey. 1982. An assessment of the degradation of agricultural lands in Canada. LRRI Contr. #118, Agriculture Canada, Research Branch, Ottawa.

Colaco, W. 1979. Denitrification in soils under a wheat-fallow rotation. M.Sc. Thesis, Dept. of Soil Science, Univ. of Saskatchewan, Saskatoon.

Dasog, G. S. 1986. Properties, genesis and classification of clay soils in Saskatchewan. Ph.D. Thesis, Dept. of Soil Science, Univ. of Saskatchewan, Saskatoon.

de Jong, E. and E. H. Halstead. 1987. Innovative Acres 1986 Report. Publ. M80, The Saskatchewan Institute of Pedology, University of Saskatchewan, Saskatoon.

de Jong, E. and K. B. MacDonald. 1975. The soil moisture regime under native grassland. Geoderma 14:207-221.

de Jong, E. and J. W. B. Stewart. 1973. Description of site: III. Soil characterization. Tech. Rep. 36, Matador Project, Canadian Committee for the International Biological Programme. Univ. of Saskatchewan, Saskatoon.

de Jong, R. 1988. Comparison of two soil-water models under semi-arid growing conditions. Can. J. Soil Sci. 68:17-27.

de Jong, R. and A. Bootsma. 1988. Estimated long-term soil moisture variability on the Canadian Prairies. Can. J. Soil Sci. 68:307-321.

de Jong, R. and H. N. Hayhoe. 1984. Diffusion-based soil water simulation for native grassland. Agric. Water Manage. 9:47-60.

Dyer, J. A., W. Baier, H. N. Hayhoe and G. Fisher. 1978. Spring field work day probabilities for selected sites across Canada. Tech. Bull. 86, Agriculture Canada, Research Branch, Ottawa.

Evans, R. 1980. Mechanics of water erosion. In M. J. Kirkby and R. P. C. Morgan (ed) Soil Erosion. John Wiley & Sons, Chichester.

Gardner, W. R. 1960. Soil water relations in arid and semi-arid conditions. In Plant-Water Relationships in Arid and Semi-arid Conditions. Arid Zone Research XV:37-61.

Geiger, R. 1959. The climate near the ground. Harvard University Press.

Gerard, C. J., P. Sexton and G. Shaw. 1982. Physical factors influencing soil strength and root growth. Agron. J. 74:875-879.

Gray, D. M., D. I. Norum and R. J. Granger. 1984. The prairie soil moisture regime: Fall to seeding. p. 159-207. In G. W. Hass (ed) The Optimum Tillage Challenge. Division of Extension and Community Relations, Univ. of Saskatchewan, Saskatoon.

Hilliard, C. R., H. P. W. Rostad and R. D. Stushnoff. 1988. Wind erodibility of Saskatchewan soils. Report of 1987-88 results. Publ. M87, Saskatchewan Institute of Pedology, Univ. of Saskatchewan, Saskatoon.

Hinman, W. C. and F. Bisal. 1968. Alterations of soil structure upon freezing and thawing and subsequent drying. Can. J. Soil Sci. 48:193-197.

Hodgson, A. S. and D. A. MacLeod. 1989. Use of oxygen flux density to estimate critical air-filled porosity of a Vertisol. Soil Sci. Soc. Am. J. 53:355-361.

Johnson, W. C. 1962. Controlled soil cracking as a possible means of moisture conservation on wheatlands of the southwestern Great Plains. Agron. J. 54:323-325.

Johnson, W. E. 1983. Soil conservation: Canadian prairies. In H. E. Dregne and W.O. Willis (ed) Dryland Agriculture. Agron. 23:259-272.

Johnston, J. R. and H. O. Hill. 1945. A study of the shrinkage and swelling properties of Rendzina soils. Soil Sci. Soc. Am. Proc. 9:24-29.

Kirkland, J. A. 1986. The effects of zero and conventional tillage systems on soil physical properties. M.Sc. Thesis, Dept. of Soil Science, Univ. of Saskatchewan, Saskatoon.

Mitchell, J., H. C. Moss and J. S. Clayton. 1944. Soil survey of southern Saskatchewan from Township 1 to 48 inclusive. Soil Survey Report No. 12, Fourth Printing, Extension Division, Univ. of Saskatchewan, Saskatoon.

Patterson, G. W. 1985. Effect of fall stubble management on overwinter soil moisture storage and crop yield. M.Sc. Thesis, Dept. of Soil Science, Univ. of Saskatchewan, Saskatoon.

Pierce, F. J., W. E. Larson, R. H. Dowdy and W. A. P. Graham. 1983. Productivity of soils: Assessing long-term changes due to erosion. J. Soil Water Conserv. 38:39-44.

Ripley, E. A. 1973. Description of site: II. Climatology of the Matador area. Tech. Rep. 12, Matador Project, Canadian Committee for the International Biological Programme. Univ. of Saskatchewan, Saskatoon.

Ritchie, J. T., D. E. Kissel and E. Burnett. 1972. Water movement in undisturbed swelling clay soil. Soil Sci. Soc. Am. Proc. 36:874-879.

Selim, H. M. and D. Kirkham. 1970. Soil temperature and water content changes during drying as influenced by cracks: A laboratory experiment. Soil Sci. Soc. Am. Proc. 34:565-569.

Shulgin, A.M. 1957. The temperature regime of soils. Israel Program for Scientific Translations in 1965.

Taylor, H. M., G. M. Roberson and J. J. Parker, Jr. 1966. Soil strength-root penetration relationships for medium- to coarse-textured soil materials. Soil Sci. 102:18-22.

van Wijk, W. A. and D. A. de Vries. 1963. Periodic temperature variations in homogeneous soil. In W. R. van Wijk (ed) Physics of Plant Environment. North-Holland Publ. Co., Amsterdam.

Warrick, A. W. and D. R. Nielsen. 1980. Spatial variability of soil physical properties in the field. p. 319-344. In D. Hillel (ed) Applications of Soil Physics. Academic Press, New York.

Wesseling, J. 1974. Crop growth in wet soils. In J. van Schilfgaarde (ed) Drainage for Agriculture. Agron. 17:7-37.

Wilson, S. J. and R. U. Cooke. 1980. Wind erosion. In M. J. Kirkby and R. P. C. Morgan (ed) Soil Erosion. John Wiley & Sons, Chichester.

Yang, S. J. and E. de Jong. 1972. Effect of aerial environment and soil water potential on the transpiration and energy status of water in wheat plants. Agron. J. 64:574-578.

Alternative Classification of Soils with Aridic Soil Moisture Regimes

R.J. Engel, J.E. Witty, and J.D. Nichols*

Introduction

Soil Taxonomy, A Basic System of Soil Classification for Making and Interpreting Soil Surveys,[1] has a structure for relating soils to each other. According to Dr. Richard Arnold,[2] "Classes at the order level are separated on the basis of properties resulting from the major processes of soil formation." Properties used are, in general, those that are thought to be the result of soil genesis. Soil climate is used as order criteria only in the orders of Aridisols and Ultisols.

The objective of *Soil Taxonomy* is to provide hierarchies of classes that permit us to understand, as fully as existing knowledge permits, the relationship between soils and the factors responsible for their character (Soil Taxonomy).

The International Committee on Aridisols, ICOMID, issued circular letter version 6.0 on April 13, 1989. This is a major revision of the Aridisol order. Now, during this second major study tour of Aridisols in the United States, is an opportune time to examine of some of the basic premises of the Aridisol order.

Soil Taxonomy

The order of Aridisols is based largely on the soil having an aridic moisture regime. In addition, the key to soil orders requires Aridisols to have one of the following, whose upper boundary is within 100 cm of the soil surface: petrocalcic, calcic, gypsic, petrogypsic, or cambic horizon or a duripan, or an argillic or natric horizon and an epipedon that is not both massive and hard or very hard when dry. Some Aridisols do not require an aridic moisture regime but have a salic horizon within 75 cm of the surface and are saturated with water within 100 cm of the surface for 1 month or more in some years and do not have an argillic or natric horizon.

The Aridisol order is the only order to have the soil moisture regime diagnostic at the order level. However, the Ultisol order requires a temperature regime that is mesic, isomesic, or warmer at the order level.

A proposal to delete the temperature regime as criteria at the order level in Ultisols is being developed in New York State.[3]

*USDA, Soil Conservation Service

"Classes at the suborder level are separated within each order on the basis of soil properties that are major controls, or reflect such controls, on the current set of soil-forming processes," according to Arnold.[2] Moisture regimes are commonly used as a criteria for suborders in *Soil Taxonomy*. The moisture regime is the basis of 26 suborders and of at least one suborder in all orders except Aridisols.

Soils with an aridic moisture regime are not treated consistently in Soil Taxonomy. With the approval of National Soils Taxonomy Handbook notice no. 13, 223 subgroups in Soil Taxonomy either are required to have or are permitted to have an aridic moisture regime. The order of Aridisols includes 137, or 61 percent, of these subgroups.

The remaining 86, or 39 percent, of the subgroups are in the orders of Alfisols, Andisols, Entisols, Mollisols, Oxisols, and Vertisols. In addition, Xeralfs are soils with an aridic moisture regime bordering on a xeric moisture regime and having both an argillic or natric horizon and a surface layer that is both hard or very hard and is massive. Xeralfs are not included in these numbers. Table 1 lists subgroups in orders other than Aridisols that are permitted to have or are required to have an aridic moisture regime.

Soils with an aridic moisture regime are treated, like other moisture regimes, at the suborder level in the orders of Andisols, Oxisols, and Vertisols. They are treated, like other moisture regimes, at the great group level, in the order of Entisols. Soils with an aridic moisture regime are recognized at the subgroup level in the order of Mollisols and, where recognized above the series level, in the order of Alfisols.

This paper examines some of the consequences of relegating the aridic soil moisture regime to a suborder or lower level and of eliminating the Aridisol order. Such an examination

[1]Soil Survey Staff, "Soil Taxonomy, a Basic System of Soil Classification for Making and Interpreting Soil Surveys." USDA Agric. Handbook No. 436, U.S. Govt. Print. Off., Washington, D.C. 1975.

[2]Arnold, Richard W. "Soil Taxonomy, A Tool of Soil Survey."

[3]Hanna, Willis E. Personal Communications. July, 1989.

Table 1: Subgroups Permitted Or Required To Have An Aridic Moisture Regime In Orders Other Than Aridisols

ALFISOLS
- Haplustalfs
  - Aridic Haplustalfs
- Kandiustalfs
  - Aridic Kandiustalfs
- Kanhaplustalfs
  - Aridic Kanhaplustalfs
- Paleustalfs
  - Aridic Paleustalfs
  - Calciorthidic Paleustalfs
- Haploxeralfs
- Palexeralfs

ANDISOLS
- Vitritorrands
  - Lithic Vitritorrands
  - Petrocalcic Vitritorrands
  - Duric Vitritorrands
  - Aquic Vitritorrands
  - Calcic Vitritorrands
  - Typic Vitritorrands

ENTISOLS
- Torrifluvents
  - Typic Torrifluvents
  - Anthropic Torrifluvents
  - Durorthidic Torrifluvents
  - Durorthidic Xeric Torrifluvents
  - Ustertic Torrifluvents
  - Ustic Torrifluvents
  - Vertic Torrifluvents
  - Xeric Torrifluvents
- Torriorthents
  - Typic Torriorthents
  - Aquic Torriorthents
  - Aquic Durorthidic Torriorthents
  - Durorthidic Torriorthents
  - Durorthidic Xeric Torriorthents
  - Lithic Torriorthents
  - Lithic Ustic Torriorthents
  - Lithic Xeric Torriorthents
  - Ustertic Torriorthents
  - Ustic Torriorthents
  - Vertic Torriorthents
  - Xerertic Torriorthents
  - Xeric Torriorthents
- Torripsamments
  - Typic Torripsamments
  - Durorthidic Xeric Torripsamments
  - Lithic Torripsamments
  - Ustic Torripsamments
  - Xeric Torripsamments

MOLLISOLS
- Argiborolls
  - Abruptic Aridic Argiborolls
  - Aridic Argiborolls
- Calciborolls
  - Aridic Calciborolls
- Haploborolls
  - Aridic Haploborolls
  - Torrifluventic Haploborolls
  - Torriorthentic Haploborolls
- Natriborolls
  - Aridic Natriborolls
- Argiustolls
  - Aridic Argiustolls
  - Torrertic Argiustolls
- Calciustolls
  - Aridic Calciustolls
  - Torrertic Calciustolls
- Durustolls
  - Aridic Durustolls
  - Orthidic Durustolls
- Haplustolls
  - Aridic Haplustolls
  - Salorthidic Haplustolls
  - Torrertic Haplustolls
  - Torrifluventic Haplustolls
  - Torriorthentic Haplustolls
  - Torroxic Haplustolls
- Natrustolls
  - Aridic Natrustolls
- Paleustolls
  - Aridic Paleustolls
  - Calciorthidic Paleustolls
  - Torrertic Paleustolls
- Argixerolls
  - Aridic Argixerolls
  - Aridic Calcic Argixerolls
  - Durargidic Argixerolls
- Calcixerolls
  - Aridic Calcixerolls
- Durixerolls
  - Aridic Durixerolls
  - Orthidic Durixerolls
- Haploxerolls
  - Aridic Haploxerolls
  - Aridic Duric Haploxerolls
  - Calciorthidic Haploxerolls
  - Torrertic Haploxerolls
  - Torrifluventic Haploxerolls
  - Torriorthentic Haploxerolls
  - Torripsammentic Haploxerolls
- Natrixerolls
  - Aridic Natrixerolls
- Palexerolls
  - Aridic Palexerolls
  - Aridic Petrocalcic Palexerolls

OXISOLS
- Acrotorrox
  - Petroferric Acrotorrox
  - Lithic Acrotorrox
  - Typic Acrotorrox
- Eutrotorrox
  - Petroferric Eutrotorrox
  - Lithic Eutrotorrox
  - Typic Eutrotorrox
- Haplotorrox
  - Petroferric Haplotorrox
  - Lithic Haplotorrox
  - Typic Haplotorrox

ULTISOLS

VERTISOLS
- Torrerts
  - Typic Torrerts
  - Mollic Torrerts
  - Paleustollic Torrerts

seems prudent and does not necessarily indicate that the authors are advocating the change at this time. Major changes are necessarily evaluated, so that any changes made are in the best interest of soil science.

Delete Aridisol Order

If the moisture regime criteria of Aridisols were not considered at the order level, Aridisols would key out as Alfisols and Inceptisols. The two suborders of the order of Aridisols, Argids and Orthids, could be handled at a suborder level in the orders of Alfisols and Inceptisols respectively. The suborder Argids could be considered a new suborder of Alfisols, "Torralfs," and the suborder, Orthids, could be considered a new suborder of Inceptisols, "Torrepts."

Considering the aridic moisture regime at the suborder level in the orders of Alfisols and Inceptisols would make *Soil Taxonomy* more consistent. The Aridic moisture regime would then be at the same level as it is presently in the orders of Andisols, Oxisols, and Vertisols. The deletion of the Aridisol order would make all of the orders in *Soil Taxonomy* based on the presence or absence of soil properties that are assumed to reflect soil forming processes.

Duripans and calcic, gypsic, natric, and salic horizons would remain at the great group level in the suborders of "Torralfs" and "Torrepts." This is consistent with the level these horizons and features are given in other soil orders.

The great groups of the suborders, Torralfs and Torrepts, could remain the same as the great groups now in the suborders of Aridisols in *Soil Taxonomy* or could be changed as agreed upon by the ICOMID committee.

The subgroups of the great groups of Torralfs and Torrepts, with the exception of those that intergrade to the order of Alfisols and Inceptisols, respectively, also require no changes. The subgroups also could be changed as agreed upon. Table 2 lists the great groups and subgroups of the proposed suborders of Torralfs and Torrepts.

Broaden Aridisol Order

If the proposal to delete the order of Aridisols is not acceptable to the committee, a proposal to bring all soils with an aridic moisture regime and diagnostic subsurface horizons into the Aridisol order could be made.

Table 2: Proposed Great Groups and Subgroups of the Proposed Suborders of Torralfs and Torrepts

TORRALFS

- Duritorralfs
  - (aquic) Duritorralfs
  - Abruptic Xerollic Duritorralfs
  - Abruptic Duritorralfs
  - Haploxerollic Duritorralfs
  - Haplic Duritorralfs
  - Xerollic Duritorralfs
  - (Ustic) Duritorralfs
  - Typic Duritorralfs
- Haplotorralfs
  - Borollic Lithic Haplotorralfs
  - Lithic Ruptic-Entic Xerollic Haplotorralfs
  - Lithic Xerollic Haplotorralfs
  - Lithic Ustollic Haplotorralfs
  - Lithic Haplotorralfs
  - Borollic Vertic Haplotorralfs
  - Borollic Haplotorralfs
  - Xerertic Haplotorralfs
  - Ustertic Haplotorralfs
  - Vertic Haplotorralfs
  - Aquic Haplotorralfs
  - Arenic Ustollic Haplotorralfs
  - Arenic Ustic Haplotorralfs
  - Arenic Haplotorralfs
  - Durixerollic Haplotorralfs
  - Duric Haplotorralfs
  - Xerollic Haplotorralfs
  - Xeric Haplotorralfs
  - Ustollic Haplotorralfs
  - Ustic Haplotorralfs
  - (Ruptic-Entic) Haplotorralfs
  - Typic Haplotorralfs
- Nadurtorralfs
  - Aquic Haplic Nadurtorralfs
  - Aquic Nadurtorralfs
  - Haploxerollic Nadurtorralfs
  - Haplic Nadurtorralfs
  - Xerollic Nadurtorralfs
  - Typic Nadurtorralfs
- Natritorralfs
  - Lithic Xerollic Natritorralfs
  - Lithic Natritorralfs
  - Borollic Glossic Natritorralfs
  - Borollic Natritorralfs
  - Aquic Natritorralfs
  - Durixerollic Natritorralfs
  - Duric Natritorralfs
  - Glossic Ustollic Natritorralfs
  - Haplustollic Natritorralfs
  - Haplic Natritorralfs
  - Xerollic Natritorralfs
  - Ustollic Natritorralfs
  - (Glossic) Natritorralfs
  - Typic Natritorralfs
- Paletorralfs
  - Borollic Vertic Paletorralfs
  - Borollic Paletorralfs
  - (Vertic) Paletorralfs
  - Petrocalcic Xerollic Paletorralfs
  - Petrocalcic Ustollic Paletorralfs
  - Petrocalcic Ustalfic Paletorralfs
  - Petrocalcic Paletorralfs
  - (Duric) Paletorralfs
  - Xerollic Paletorralfs
  - Ustollic Paletorralfs
  - Xeric Paletorralfs
  - (Ustalfic) Paletorralfs
  - Typic Paletorralfs

TORREPTS

- Calcitorrepts
  - Borollic Lithic Calcitorrepts
  - Borollic Calcitorrepts
  - Lithic Xerollic Calcitorrepts
  - Lithic Ustollic Calcitorrepts
  - Lithic Calcitorrepts
  - Aquic Duric Calcitorrepts
  - Aquic Calcitorrepts
  - Durixerollic Calcitorrepts
  - (Duric) Calcitorrepts
  - Xerollic Calcitorrepts
  - Ustollic Calcitorrepts
  - (Xeric) Calcitorrepts
  - Ustic Calcitorrepts
  - Argic Calcitorrepts
  - Typic Calcitorrepts
- Haplotorrepts
  - Borollic Lithic Haplotorrepts
  - Durixerollic Lithic Haplotorrepts
  - Lithic Xerollic Haplotorrepts
  - Lithic Haplotorrepts
  - Natric Haplotorrepts
  - Borollic Vertic Haplotorrepts
  - Borollic Haplotorrepts
  - Xerertic Haplotorrepts
  - Ustertic Haplotorrepts
  - Vertic Haplotorrepts
  - Aquic Duric Haplotorrepts
  - Aquic Haplotorrepts
  - Durixerollic Haplotorrepts
  - Duric Haplotorrepts
  - Fluventic Haplotorrepts
  - Anthropic Haplotorrepts
  - Xerollic Haplotorrepts
  - Ustollic Haplotorrepts
  - (Xeric) Haplotorrepts
  - Ustic Haplotorrepts
  - Typic Haplotorrepts
- Duritorrepts
  - Aquentic Duritorrepts
  - Aquic Duritorrepts
  - Haploxerollic Duritorrepts
  - (Haplustollic) Duritorrepts
  - Xerollic Duritorrepts
  - (Ustollic) Duritorrepts
  - (Ustic) Duritorrepts
  - Typic Duritorrepts
- Gypsitorrepts
  - Petrogypsic Gypsitorrepts
  - Calcic Gypsitorrepts
  - Cambic Gypsitorrepts
  - Typic Gypsitorrepts
- Paletorrepts
  - (Borollic) Paletorrepts
  - Aquic Paletorrepts
  - Xerollic Paletorrepts
  - Ustollic Paletorrepts
  - (Xeric) Paletorrepts
  - Ustic Paletorrepts
  - Typic Paletorrepts
- Salitorrepts
  - Aquollic Salitorrepts
  - (Haplic) Saltorrepts
  - Typic Salitorrepts

The key to the orders proposed by ICOMID on April 13, 1989, allows Aridisols to have a mollic epipedon. This change would cause all aridic Xerolls,with an argillic, cambic, calcic, petrocalcic, gypsic, petrogypsic, natric, or a salic horizon or a duripan within 100cm, and many aridic Borolls and Ustolls, to be reclassified as Aridisols. This would affect about 16 of the 88 subgroups with an aridic moisture regime in orders other than Aridisols.

Alfisols with an aridic moisture regime would classify as Argids if the requirement that the epipedon of an Aridisol, with an argillic or natric horizon, can not be both massive and hard, or very hard when dry, were removed from the Key to Soil Orders of *Soil Taxonomy*.

The remaining subgroups with an aridic moisture regime also could be brought into the order of Aridisols by placing Aridisols above Andisols, Oxisols, and Vertisols in the key to the orders or by not allowing an aridic moisture regime in these orders. The suborders "Andids," "Oxids," and "Vertids" could be added to the Aridisol order for those soils with an aridic moisture regime now in the orders of Andisols, Oxisols, and Vertisols, respectively.

Summary

In summary, classes at the order level are, in most cases, separated on the bases of properties that resulted from the major soil-forming processes. The order Aridisols does not follow this principle. The order of Aridisols could be treated as suborders of Alfisols and Inceptisols with very little change to *Soil Taxonomy*.

If the order of Aridisols is not deleted, a proposal could be made to bring all soils with an aridic moisture regime and diagnostic subsurface horizons into the order of Aridisols.

References

Arnold, Richard W. "Soil Taxonomy, A Tool of Soil Survey." International Committee on Aridisols, "Aridisols" Draft: April 13, 1989. Unpublished.

Soil Survey Staff. "National Soil Taxonomy Handbook," 430-VI, Issue No. 13. USDA, Soil Conservation Service, Washington, D.C., 1989.

Soil Survey Staff. "Soil Taxonomy, a Basic System of Soil Classification for Making Soil Surveys." USDA Agric. Handbook No. 436, U.S. Govt. Print. Off., Washington, D.C., 1975.

Witty, John E., and Richard W. Arnold. "Soil Taxonomy, an Overview." Outlook on Agriculture, Volume 16, No. 1, 1987, Copyright 1987. Pergamon Journals Ltd. Printed in Great Britain.

Soil Forming Processes in Soils with Cryic and Frigid Soil Temperature Regimes in Idaho

M.A. Fosberg, A.L. Falen, R.R. Blank, and K.W. Hipple[1]

Abstract

Idaho has extensive areas of Aridisols with cryic and frigid soil temperature regimes. The pedogenesis of these soils is controlled by landforms and parent materials, periglacial activity and patterned ground formation, climate, and plant community composition.

Aridisols with cryic soil temperatures occur at high elevations mostly in east-central Idaho and developed principally in limestone alluvium/colluvium with variable amounts of calcareous loess. Periglacial climate and patterned ground formations have strongly affected these soils, causing, among other changes, pronounced particle size discontinuities in the soil profile and spatial dependence of pedon properties at a very short scale. Duric horizons characterize pedons older than late-Pliestocene.

Aridisols with frigid soil temperatures occur at lower elevations in east-central and southeastern Idaho and on the Owyhee Plateau of southeastern Idaho. Most formed in calcareous loess. Those that occur in eastern Idaho typically have A, Bw, Bk, and C genetic horizons and those on the Owyhee Plateau have A, Bt, Bk, and Bkqm genetic horizons.

Introduction

Idaho has extensive areas of Aridisols with frigid and cryic soil temperature regimes. They occur at elevations of 1400 to 1700 m and receive 170 to 300 mm precipitation. Detailed climatic information is given by Hipple et al. (1987) for the cryic soils. They also occur in other western states, such as Wyoming, Utah, and Nevada, that have high elevations and similar climatic characteristics. In Idaho, they occur primarily in the east central part, from Salmon to Idaho Falls, and on the Owhyee Plateau in southeastern Idaho (Fig. 1) (Fosberg and McGrath, 1988; USDA-SCS, 1984). The acreage summary from the soil moisture and temperature map by Fosberg and McGrath (1988) gives a total of 142,777 ha (352,803 acres) of Aridisols with cryic soil temperatures and 1,495,593 ha (3,695,611 acres) of Aridisols with frigid soil temperatures in Idaho.

This paper discusses the factors and processes influencing the development and characteristics of cold Aridisols in Idaho. Most of the discussion concerns soils with cryic soil temperature regimes with some references to soils with frigid soil temperature regimes. The topics of discussion are: (1) general morphological and physical properties of selected soil series; (2) landform and parent material relationships; (3) effects of periglacial climate and patterned ground on soil genesis; (4) characteristics and development of duripans; (5) chemical and physical properties of diagnostic horizons, habitat type communities, and soil climate relationships. The conclusions proposed are a synthesis of research not only by the authors but also by many other scientists.

Results and Discussion

General Morphology of Soils

Aridisols with Cryic Soil Temperature Regimes

Most cryic soils in Idaho formed in carbonate rich parent material derived from limestones, carbonatic mudstones and siltstones, and smaller amounts of calcareous loess. Several genetic horizon sequences represent the cryic soils (see Table 1).

These soils all have similar solums with a thin calcareous A horizon and weakly developed Bk horizons. Differences among pedons occur in the degree of cementation, depth to cementation, and coarse fragment content of the lower Bkq, Bqk, and Bqkm horizons. The texture in surface horizons is usually gravelly loam or very gravelly loam with increasing coarse fragments and sand with depth. Table 1 gives pH, organic carbon (O.C.), calcium carbonate equivalent (CCE), coarse fragment, and textural class with modifier for 5 representative cryic soils.

[1]M.A. Fosberg and A.L. Falen, Div. of Soils, Dept. of Plant, Soil, & Ento., University of Idaho, Moscow, ID; R.R. Blank, USDA-ARS, Reno, NV; K.W. Hipple, USDA-SCS State Office, Spokane, WA. Contribution from the College of Agriculture, University of Idaho, Research Paper No. 89759, Received 19 Dec. 1989.

Aridisols with Frigid Soil Temperature Regimes

These Aridisols in the upper Snake River Plain are weakly developed. The soil surveys of this area (USDA-SCS, 1973, 1979, 1981) do not describe any cemented horizons in these soils. Table 2 gives typical genetic horizon sequences, plus pH, O.C., CCE, and textural properties representing one soil from this area. The Aridisols formed in calcareous loess or calcareous sand and gravelly alluvium or in a mixture of these materials. They all are relatively weakly developed with thin A and B horizons and are calcareous to the surface. Soils formed in deep loess, however, have an A, Bw, and Bk horizonation sequence. Extensive areas of these soils in the northern part of the Snake River Plain have been reworked by wind to form low incipient sand dunes aligned in a southwest to northeast direction.

The Aridisols with frigid soil temperature regimes, on the Owhyee Plateau in southwest Idaho, are represented by a typical horizon sequence of A-Bt-Bkq-2Bkqm. These soils are on very old landscapes. They have developed strong textural B horizons and strongly cemented silica duripans. Many have been influenced by periglacial climates and solifluction.

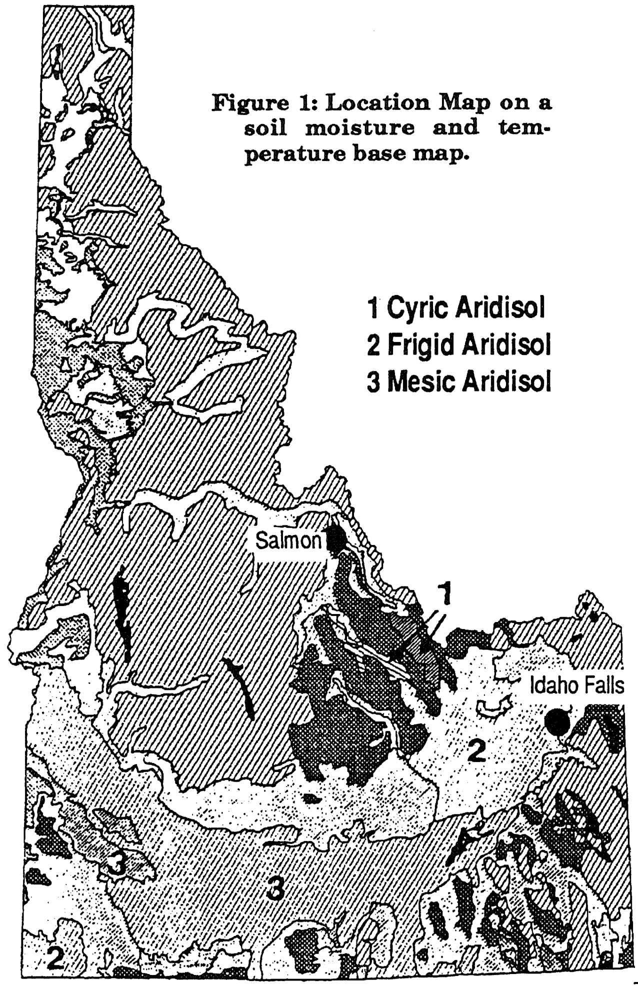


Figure 1: Location Map on a soil moisture and temperature base map.

Landform and Parent Materials

Soils With Cryic Temperatures

The rugged basin and range topography in east-central Idaho, where Aridisols with cryic soil temperature regimes occur (Fig. 1), is a portion of the Basin and Range Province which joins the north side of the Snake River Plain. In Fig. 1, the major cryic soil area is designated as 1. It is composed of four northwest to southeast aligned mountain ranges and three broad river valleys. Along the east boundary is the Beaverhead Mountains of the Bitterroot Mountain Range (Continental Divide). The Lemhi and Lost River Mountain Range occupy the central ranges, where north to south, the Salmon, Sheep, and Whiteknob Mountains border on the west side of this area. The broad valleys where the Aridisols with cryic soil temperature regimes occur are occupied from east to west by the Lemhi, Pahsimeroi, and Salmon Rivers in the north and the Birch Creek, Little Lost, and Big Lost Rivers in the south.

Fan terrace formation and glaciation

These valleys are made up of extensive fan terrace systems characterized by transported sediment eroded from the adjacent mountainous uplands. The sediment deposited in the fan terraces is primarily poorly sorted carbonate gravels derived from carboniferous limestones and some Devonian and Silurian dolomite rocks (Funk, 1977).

A study by Funk (1977) on the formation of fans terraces in Birch Creek Valley and studies of glaciation in the adjacent mountains by Knoll (1973) and Dort (1969) place fan terrace system development during the Quaternary glacial chronologies of the area. The two major glacial episodes documented for this area are the Bull Lake and Pinedale. They have been dated by obsidian hydration by Pierce et al. (1976) at 140,000 to 70,000 years B.P. and 70,000 to 10,000 years B.P., respectively. The Bull Lake correlates with the late Illinoian and Sangamon and the Pinedale with the Wisconsin stages of the midcontinent glaciations.

There also were pre-Bull Lake glacial episodes (unnamed) and post-Pinedale episodes of Temple Lake (10,000 to 3000 yrs. B.P.) and

Neoglaciation (3000 to 1000 yrs. B.P.). Evidence for these glacial episodes is found in the moraine deposits at the mouths of canyons below glacial cirque formations of the mountain ranges.

Fan terrace characteristics

The configuration of the fan terrace system usually consists of several inset fan segments. This indicates that the overall gradient of each fan system has decreased with time. Studies by Fosberg, in cooperation with Funk (1977), show a correlation of increasing silica and carbonate cementation and development of the soil duripans with increasing age of alluvial fan terraces. This cementation ranged from pendants on coarse fragments, to very thin (1 mm) laminar caps, to discontinuous cementing in the younger soils and fans, to continuous thick indurated cementation up to 30 cm thick in the soils on older fan terraces. The degree of soil development above the Bqk and Bqkm horizons was generally weak regardless of age of the fan terraces.

Soil parent material

The parent materials for most of these soils are the carbonate rich alluvium and the carbonate rich loess from the Snake River Plain.

Table 1. Representative Aridisol soil series with cryic soil temperature regimes and their genetic horizon sequence.

Horizon	Depth cm	pH[1]	O.C.[2] %	CCE[3] %	>2mm %	Texture
Unnamed (mound) v. gr. loam: Loamy-skeletal, mixed, frigid Xerollic Calciorthids -h.t.-*A. wyomingensis / Agropyron spicatum*						
A	0-5	7.6	2.82	31	56	GRV - L
Bk1	5-15	7.9	1.96	32	43	GR - L
Bk2	15-42	8.0	1.69	35	38	GR - L
Bk3	42-61	7.6	1.61	36	58	GR - L
Bkq	61-100	7.8	0.90	39	62	GRV - L
2Bqk	100-140	8.3	0.13	70	86	GRX - SL
Leatherman v. gr. loam: Loamy-skeletal, carbonatic, frigid, shallow Xerollic Durorthids - h.t.-*A. arbuscula / Agropyron spicatum*						
A	0-13	6.8	5.00	31	21	GRV - L
Bk	13-28	7.4	1.74	39	32	GRV - L
2Bqkm	28-40	7.7	0.54	80	86	Indurated duripan
2Bqk	40-150	7.9	0.13	82	95	GRX - LS
Arbus (intermound) gr. loam: Sandy-skeletal, carbonatic, frigid Xerollic Calciorthids - h.t.-*A. arbuscula / Agropyron spicatum*						
A	0-10	7.6	2.13	2	47	GR - L
Bk	10-23	7.6	1.63	36	63	GRV - L
Bkq1	23-33	7.9	0.88	66	84	GRV - SL
Bkq2	33-43	8.4	0.69	70	84	GRV - SL
Bkq3	43-76	8.6	0.16	76	88	GRX - LCOS
Bkq4	76-152	8.6	0.11	71	86	GRX - COS
Bluedome loam: Coarse-loamy, carbonatic, frigid Xerollic Durorthids - h.t.-*A. arbuscula / Agropyron spicatum*						
A	0-8	7.1	3.22	6	11	L
Bk1	8-28	7.3	2.22	17	17	L
Bk2	28 43	7.3	1.48	39	14	L
Bk3	43-85	7.2	1.24	58	64	L
2Bqkm	85-115	7.4	1.48	55	232	Indurated duripan
2Bqk1	115-133	7.7	0.69	55	80	GRX - COS
2Bqk2	133-150	7.8	0.45	55	79	GRX - COS
Nitchly gr. silt loam: Loamy-skeletal, carbonatic Xerollic Calciorthids - h.t.-*A. nova / Agropyron spicatum*						
A	0-10	7.3	2.98	15	39	GR - SIL
Bk1	10-18	7.3	1.83	80	45	GR - SIL
Bk2	18-33	7.2	0.78	41	58	GRV - L
Bk3	33-58	7.6	0.37	73	62	GRV - L
Bkq1	58-84	7.5	0.38	54	58	GRV - CL
Bkq2	84-107	7.4	0.46	88	78	GRX - CL
Bkq3	107-152	7.5	0.23	62	89	GRX - CL

[1]pH determined on saturated paste.
[2]O.C. determined by $K_2Cr_2O_7$ digest and $FeSO_4$ titration.
[3]CCE determined by acid neutralization and subsequent back titration with a standardized base.

Soils With Frigid Temperatures

The major areas of Aridisols with frigid soil temperature regimes occur in the large area joining the south side of Aridisols with cryic soil temperature regimes (in Fig. 1, designated as 2), along the higher elevation fringes of the broad extensive area of mesic Xerollic Aridisols on the Snake River Plain that are designated by 3, and on the northern fringes of the Owhyee High Plateau (Harkness, 1983).

Landform and age

The large area of Aridisols with frigid soil temperature regimes, at the east end of the Snake River Plain, in the area west and north of Idaho Falls and Blackfoot, are on a broad plain made up of a series of Pliocene and Pliestocene basalt flows, some as young as 2,000 and 4,000 years B.P. Older basalt flows are capped with calcareous loess. On the north half, large areas of calcareous alluvium, coming from the Big Lost, Little Lost, and Birch Creek river valleys to the north, cover the older basalt flows. The Owyheee Plateau is dominated by welded tuff volcanics of the Miocene age.

Soil parent material and age

The loess parent materials in south Idaho are extensive (Pierce et al., 1982; Lewis and Fosberg, 1982; Blank, 1987). In southeastern Idaho, loess layers were dated by Pierce et al. (1976 and 1982) and correspond to the Bull Lake and Pinedale glacial episodes described in the previous section for soils with cryic temperatures. Pinedale loess is older than 15,000 yrs. B.P. because the Bonneville flood, dated about 15,000 yrs. B.P., has scoured the loess along the Sanke River. The major source of this loess is attributed to the glacial outwash flood deposits of the Snake and other rivers.

Effects of Periglacial Climates and Patterned Ground Formation

Patterned ground is a term for the more or less symetrical forms such as circles, polygons, nets, stripes, garlands, and steps that are charactersitic of, but not confined to, mantles sub-

jected to intense frost action in periglacial areas (USDA-SCS, 1986). Patterned ground has been observed by the authors throughout Idaho and other western states and indeed occurs throughout the world (Krantz et al. 1988). It is a relict of former periglacial climates. It is best developed in close proximity to areas of glaciation at higher elevations.

The most recent theory on the geophysical process in their formation is by Ray et al. (1983). Two excellent monographs (Price, 1972, and Washburn, 1980) review the extensive literature on patterned ground. In Idaho, three studies (Fosberg and Hironaka, 1964; Fosberg, 1965; Malde, 1964) show the relationship of patterned ground to (1) its origin from effects of cold periglacial climates, (2) the characteristics and genesis of soils, and (3) the effects of these soil properties on the distribution of sagebrush habitat types.

Soil Variability and Periglacial Climates

In the Aridisol area with cryic temperatures within the boundaries of a common landform, it appears that the effects of former periglacial climates and patterned ground formation account for much of the soil variability, as illustrated by the representatives in Table 1. Potential variations in pedon properties that could result from the effect of periglacial climates are: variations of Bk verses Bkq or Bqk horizon sequences; variable degree of cementation; differences in depth to indurated duripans; thickness of the solum; and the differences in particle-size distribution and gravel content in the A and B horizons.

Areas with No Periglacial Influence

In Idaho, Oregon, and Washington a major criteria necessary for patterned ground formation is a restrictive layer (hardpan, claypan, bedrock) at shallow depths. In the major frigid area of the eastern Snake River Plain, many soils have developed in late Pliestocene or Holocene deep loess alluvium. Insufficient time for the formation of argillic or cemented pans precludes the production of patterned ground.

Areas With Periglacial Influence

In the Owyhee Plateau these soils are strongly developed, having argillic and duripan horizons, and are often shallow to bedrock. Patterned ground is a common feature.

Table 2. Representative Aridisol soil series with frigid temperature regimes and their genetic horizon sequence.

Horizon	Depth cm	pH[1]	O.C.[2] %	CCE[3] %	>2mm %	Texture
Pancheri silt loam: Coarse-silty, mixed, frigid Xerollic Calciorthids - h.t.-*A. wyomingensis / Stipa thurberiana*						
A1	0- 5	6.8	2.14	0	TR	SIL
A2	5-10	-	1.42	0	0	SIL
Bw	10-30	-	0.98	0	0	SIL
Bk1	30-43	7.7	0.79	7	0	SIL
Bk2	43- 71	8.4	0.31	24	0	SIL
Bk3	71-96	8.7	0.19	21	0	SIL
Bk4	96-112	8.7	0.22	19	0	SI
C	112-162	8.5	0.11	18	0	SI
Grassy Butte loamy sand: Sandy, mixed, frigid Typic Calciorthids - h.t. - *A. wyomingensis / Stipa comata*						
A	0-18					LS
Bk1	18-48					LS
Bk2	48-88					LS
C1	88-128					LS
C2	128-150					LS
Malm stony sandy loam: Coarse-loamy, mixed, frigid Xerollic Calciorthids - h.t. - *A. wyomingensis / Stipa comata*						
A	0-10					ST - SL
Bk	10-30					SL
C	30-60					SL
2R	60+					Basalt
Whiteknob gravelly loam: Sandy-skeletal, mixed, frigid Xerollic Calciorthids - h.t.-*A. wyomingensis / Agropyron spicatum*						
A	0-13					GR - L
Bk1	13- 23					GR - L
Bk2	23- 35					GR - L
Bk3	35- 58					GRV - SL
2C	58-150					GRV - S
Heckison silt loam: Fine-loamy, mixed, frigid Xerollic Durargid. - h.t.-*A. wyomingensis / A. spicatum*						
A	0-10	6.5	1.22	0	9	SIL
Bt1	10-23	6.6	1.09	0	7	SICL
Bt2	23-33	7.2	0.66	0	13	SICL
Bk	33-48	8.0	0.77	26	18	SIL
Bkq	48-79	7.9	0.60	33	61	GRV - SIL
2Bkqm	79-97	-	0.56	66	92	GRX - SIL

[1]pH determined on saturated paste.
[2]O.C. determined by $K_2Cr_2O_7$ digest and $FeSO_4$ titration.
[3]CCE determined by acid neutralization and subsequent back titration with a standardized base.

Characteristics and Development of Duripans

Duripans Intergrading with Petrocalcics

General appearance

Soils with frigid and cryic temperatures occurring in the horst and graben topography of east-central Idaho often contain subsurface cemented horizons. These pan-containing soils are dominantly located on fan terraces derived from limestones. Cemented regions consist of a clast-supported framework of rounded gravels and cobbles with variable amounts of matrix material. Pans are laterally continuous and range from 10 to 50 cm in thickness. Maximum moisture penetration into the soil profile likely controls the depth to the cemented layers as well as the parent material discontinuities.

The appearance of cemented horizons changes with age. As time increases, pendants

become thicker and begin to bridge individual framework clasts eventually coalescing under several clasts. The laminar cap becomes thicker and more indurated. In incipient and youthful pans, there is little matrix material among framework clasts and/or the matrix material is not cemented. In mature pans, void space among framework clasts is filled largely with matrix material, and the matrix often is indurated.

Duripan criteria

The large proportion of limestone clasts, and the fact that calcium carbonate is the dominant cementing agent, requires that these pans be considered intergrades with petrocalcics. We examined many pans in regard to duripan criteria as defined by *Soil Taxonomy* (1975), with variable results. Some pan fragments completely dissolved in 1 N HCl, yet other fragments from the same sampling location remained partially intact. These findings were noted regardless of pan maturity. The residue after HCl treatment, however, always contained opaline silica. Moreover, in thin sections, opaline silica was witnessed in laminar caps and in some pendants. The silica is undoubtedly causing local cementation - bridging of particles - but much of it occurs as individual, discrete, silt-sized particles that strengthens the pans. The presence of this secondary opaline material qualifies the cemented horizons as duripans.

Nature of cementation

Three distinct regions of cementation occur in these duripans: the laminar cap, bridging of framework clasts by matrix material, and coalescing pendants.

The laminar cap consists of laminations and striations, 0.04 to 0.10 mm thick, of alternating micritic calcite with brown microgranular calcite-opaline silica material (an intimate association of calcite with opaline silica). Scanning electron microscope observations shows the opaline silica occurs in web-like arrangements that both bridges particles and occurs as isolated units. In mature duripans, framework clasts often are cemented together through the bridging of matrix material.

Most of the material dissolves with HCl treatment, indicating that the dominant cementing agent is calcium carbonate. However, in the oldest duripans studied, regions containing microgranular calcite-opaline silica similar to that in the laminar cap were seen. It is not fully understood if the calcium carbonate cement was emplaced via percolating water or if *in situ* solution-precipitation of limestone occurred in thin water films around particles to produce the cement.

It should be mentioned that a portion of the limestone is dolomitic and is, therefore, unstable in the pedogenic environment. Most dolomite clasts have a thin alteration rind of calcite. Many framework clasts have a pendant emanating from beneath (Blank and Fosberg, In press, a). At the lower boundary of the duripan these pendants often coalesce and, therefore, bridge two or more framework clasts. Calcite is the dominant component of pendants; however, thin concentric layers of opaline silica occur in some pendants. The opal creates a brownish-green coloration and adamantine luster that is evident in the field (Blank and Fosberg, In press, a).

The role of eolian dust

The eolian dust is an integral factor in the genesis of these duripans. The presence of eolian material is indicated by the large proportion of allochthonous minerals in the duripans (minerals such as certain feldspars, amphiboles, and pyroxenes) that could not have been derived from the parent fan terrace materials. The eolian dust serves as a source of silica, the mechanism of which this paper presents later. Eolian dust is a major component of the matrix material filling voids among the framework clasts.

Eolian dust also may serve as a source of calcium carbonate for cementation. This statement may seem odd considering that limestone rock is a dominant parent material in the duripan-containing soils. However, it has been observed that the crystallinity or inherent chemical composition of the limestone material makes it resistant to weathering. In dissolution experiments with HCl, limestones, even silt-sized fragments, dissolved much slower than dirty gray micritic calcite (eolian transported).

The importance of eolian dust also is indicated by the mineralogy of the clay-sized fraction. Mica (illite), kaolinite, and quartz are the major clay-sized minerals in the carbonate-free residue of limestone. However, clay-sized residue of HCl-treated whole duripan fragments often contain appreciable smectite. The smectite is brought in with eolian dust (Lewis and Fosberg, 1982).

Age and mineralogy relationships

Incipient duripans (Holocene aged) and duripans as old as late-Pliestocene dominantly ARE smectitic in the clay-sized non-carbonate fraction. In older duripans (40,000 to 140,000 years

B.P.), sepiolite and polygorskite assume importance and smectite is altered to these minerals over time.

Duripans of the Owyhee Plateau

General appearance and micromorphology

The best developed of these duripans directly overlay Miocene to Pliocene basalt on large shield volcanos. They also occur over more siliceous extrusive volcanics and on ignimbrites. Several lines of evidence suggest these duripans are distinctly different from the aforementioned duripans. Rather than occurring as continuously cemented layer, these duripans occur as thin plates 2 to 8 cm thick and 20 to 60 cm in diameter. The plates stack atop each other to create a thick indurated horizon.

Soil forming processes have completely obsured any indication of an original fabric in these duripans. The pre-duripan landscape may have consisted of a framework of basalt fragments in filled with loess. Whatever the initial conditions were, the duripans now consist of a very complex, densely indurated, laminar-concretionary fabric whose principle minerals include calcite and opaline-silica.

Polygenesis

The micromorphology of these duripans suggests a complex genetic history. The arrangement of fabric components suggests these duripans are polygenetic, the summation of several episodes of soil truncation with subsequent readdition of loess.

Three lines of evidence support this conclusion. First, zones of well-sorted loess agglomerates occur within the duripans, which could not have accumulated by any known subsurface soil pedogenic process. Convoluted laminar fabrics are seen in most thin sections of duripan plates, which are similar to the fabrics produced in arid regions by the action of lichens and cyanobacteria at the soil surface. Finally, at one sampling location, a laterally extensive air-fall tephra layer occurs in the middle of the duripan plates, which clearly was deposited when the duripan or proto-duripan was surficially exposed.

Mineralogy

Calcite is the dominant mineral which occurs in these duripans. However, opaline silica may account for nearly 20 percent by weight of the duripan. The opal is largely opal-A, except for one duripan which was dominated by opal-CT (Jones and Segnit, 1971). All the duripans examined contain sepiolite, which occurs as needles emanating from secondary concretions. The sepiolite is not well crystallized, and TEM micrographs of sepiolite laths often show opal spheres attached to the sepiolite with a gel-like substance.

Source of silica

It generally is agreed that volcanic tephra is important in the genesis of duripans in the western United States (Flack et al., 1969, 1974). The rapid weathering kinetics of amorphous volcanic glasses contribute silica to the soil solution, which then percolates downward and precipitates near the wetting front to create the duripan (Chadwick et al., 1987).

The incorporation of eolian dust into duripans and its subsequent alteration may contribute to the silification of duripans (Blank and Fosberg, In press, b). Dust is incorporated into calcareous pans via capture of loess agglomerates or by fracture fillings to form pedotubules. Once incased within the duripans, closed-system high pH conditions and increased surface Gibbs free energy, caused by the force of calcite crystallization, leads to rapid *in situ* alteration of the loess.

The stable and end products of alteration include opal-A and uncharacterized X-ray amorphous materials. The abundance of composite particles and loess pedotublules in these duripans suggest that their contribution to the silicification of pans is of equal or greater magnitude than the commonly accepted mechanism of addition via percolating Si-rich soil solutions from overlying soil horizons.

Chemical and Physical Properties of Diagnostic Horizons, Habitat Type Communities, and Climatic Relationships

Habitat Type Concept and Soils

This discussion accepts that the sagebrush species, including *Artemisia tridentata* subspecies complex, are ecologically significant, and their distribution is related primarily to moisture, temperature, depth, and soil properties that are related to soil development (Hironaka et al. 1983). The habitat type (h.t.) classification concept used here is based on climax vegetation, where certain sagebrush and understory grass species, related to the environment, are dominant.

However, it has been found (Hironaka et al., 1983) that the requirements for the sagebrush and grass species do not always coincide. Each

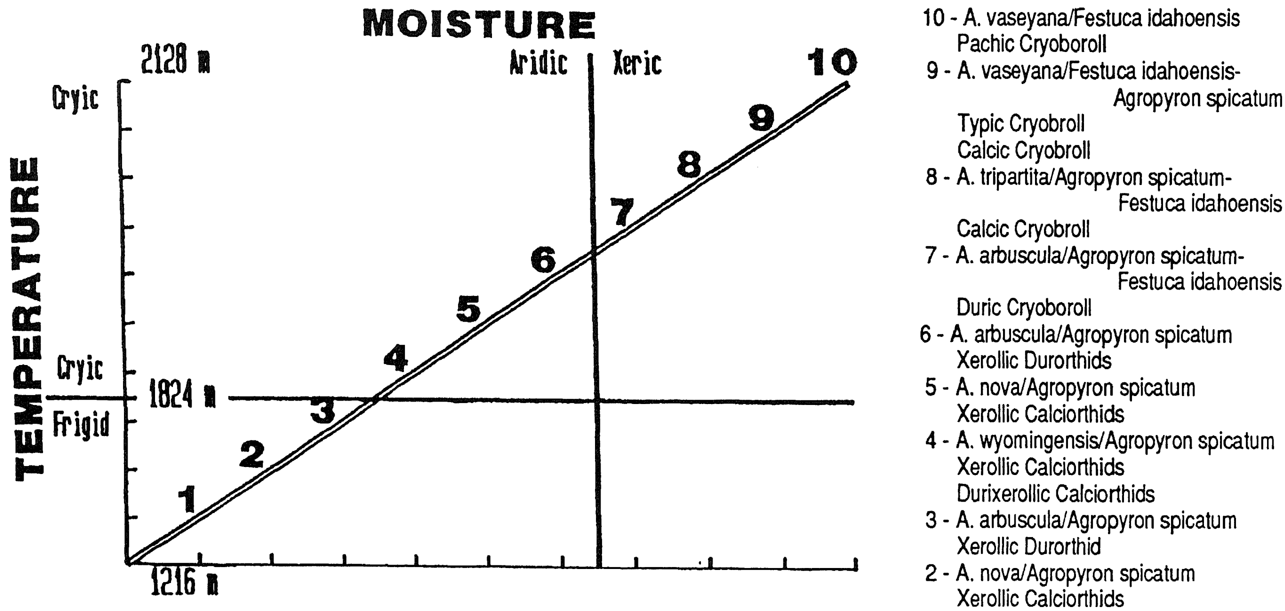


Figure 2. Ecological gradient showing moisture-temperature relationship to habitat types and soils.

will occupy certain positions of the environmental gradient. Thus, with each sagebrush-perennial grass combination, a more homogeneous environment is delineated. It holds that in each environment, the potential to produce a particular h.t. is different, such as *A. arbuscula/ Agropyron spicatum* versus *A. wyomingensis/ Agropyron spicatum*.

Figure 2 illustrates the entire ecological range found along an elevational gradient in the cryic and adjacent frigid areas. This ecological gradient shows the ecological position of each h.t. along with elevation, soil moisture, temperature, and classification of soils.

Soil Climate and Habitat Types

A research study in the Birch Creek Valley by Hauxwell (1977), in cooperation with M. Fosberg and M. Hironaka, demonstrated that as elevation increased precipitation increased. The study dealt with soil characterisitics/sagebrush h.t. relationships. Precipitation ranged from 118 mm for *A. wyomingensis* h.t. to 293 mm for *A. vaseyana* h.t. for the period of September 1975 to September 1976. Precipitation was smallest during winter and largest during summer. For all precipitation gauges, 6% of precipitation was received mid-December to March, 27% from March to mid-June, 48% from mid-June to mid-September, and 19% from mid-September to mid-December.

Hauxwell's (1977) study also determined that the date of moisture depletion to 1.5 MPa, as determined by thermocouple psychrometer readings, was, at least, by June 22 for *A. wyomingensis, A. nova,* and *A. arbuscula*, July 6 for *A. tripartita*, and July 20 for *A. vaseyana* h.t. The average summer soil temperature for all h.t. (Hauxwell's, 1977; Hipple, 1983; Hipple, 1987 were within the cryic temperature regime.

Soil climate studies by Hauxwell (1977) and Hipple (1987), ecological habitat type classification by Hironaka (1983) and an abundance of unpublished data[2] show a distinct ecological sequence for the h.t. as given in Figure 2.

Genesis of Soil Properties

Diagnostic soil properties, in addition to climate, that interrelate soil genesis and classification and effect the distribution of the plant communities are: (1) ochric epipedons and the accumulation of O.C. as it affects the placement of soils in Xerollic subgroups, (2) accumulation of calcium carbonate and silica compounds to form, respectively, calcic horizons and duripans, (3) parent material discontinuities that affects moisture percolation, and thus, duripan development, (4) particle size distribution, and (5)

[2]Field and laboratory data by M.A. Fosberg and A.L. Falen, University of Idaho, Moscow, ID; SCS soil survey for Clark, Lemhi, Custer, and Butte Counties, Idaho; SCS, National Soil Survey Laboratory, Lincoln, NE.

mound-intermound relationship to patterns of soil development.

Ochric epipedon and organic carbon increase

Progressing in an ecological sequence (Fig. 1) from Typic Aridisols supporting desert shrub h.t. (southwestern part of unit 3), through the sagebrush h.t. area in Aridisols with mesic soil temperature regimes (unit 3), through Aridisols with frigid soil temperature regimes of the Snake River Plain (unit 2), and into the cryic soils of Aridisols (unit 3), the O. C. gradually increases with decreasing temperatures (Tables 1, 2, 3). Figure 2 shows these relationships. The sagebrush-grass h.t.s associated with the soils above the Typic Aridisols are represented by Xerollic intergrades.

Many of the Aridisols with frigid soil temperature regimes, such as Pancheri (Table 2), have formed Bw or cambic horizons because they occur outside the area of periglacial influence, whereas, Aridisols with cryic soil temperature regimes have been mixed by solifluction and lack a well defined cambic or Bw horizon. In southern Idaho, cambic horizons appear to represent a younger stage of soil development because similar soils on older landforms often have argillic horizons.

Most of the Aridisols with frigid soil temperature regimes, in the upper Snake River Plain, are relatively deep and, therefore, support *A. wyomingensis*. In other areas of Aridisols with frigid soil temperature regimes, where restricting layers exist at a shallow depth, soil profiles support *A. arbuscula*.

Carbonates

The most striking feature of the Aridisols with cryic soil temperature regimes is the abundance of carbonate, which contributes to the formation of calcic horizons, to the cementation in the duripans, and the carbonatic family classification (Table 1). High amounts of carbonate in the ochric epipedons in these soils come from eolian dust and are also related to the high limestone gravel content and mixing by solifluction from periglacial climates.

In Aridisols with frigid temperatures developed in loess, carbonates have leached from the A and Bw horizons to 30 cm (Table 2). The major source for these carbonates is calcareous loess parent material. Carbonate accumulation in mesic soils (Table 3) is very comparable. Much of the carbonate source is also from the same age of Pinedale calcareous loess or the same age of sediments.

Table 3. Representative Aridisol soil series with mesic soil temperature regimes and their genetic horizon sequence.

Horizon	Depth cm	pH[1]	O.C.[2] %	CCE[3] %	>2mm %	Texture
Greenleaf silt loam: Fine-silty, mixed, mesic Xerollic Haplargids - h.t.-*A. wyomingensis / Stipa thurberiana*						
Ap	0-25	7.8	0.91	2	0	SIL
Bt1	25-31	7.9	0.38	2	0	SIL
Bt2	31-38	8.0	0.32	2	0	SIL
2Bk1	38-46	7.5	0.28	9	0	SIL
2Bk2	46-61	7.2	0.23	16	0	SIL
2Bk3	61-91	8.2	0.15	14	0	SIL
3Bk	91-152	8.2	0.16	17	0	SIL
Turbyfill very fine sandy loam: Coarse-loamy, mixed, calcareous, mesic Xerollic Calciorthids - h.t.-*A. wyomingensis / Stipa thurberiana*						
Ap	0-23	7.7	0.71	2	0	VFSL
Bk1	23-51	8.0	0.40	16	0	L
Bk2	51-69	8.0	0.26	14	0	VFSL
2C	69-89	8.1	0.13	8	5	LS
3C	89+	8.0	0.05	5	27	LS
Scism loam: Coarse-silty, mixed, mesic Haploxerollic Durorthids - h.t.-*A. wyomingensis / Stipa thurberiana*						
A1	0-4	7.4	1.49	0	0	L
A2	0-11	7.3	0.63	0	0	L
Bw	11-23	7.7	0.46	0	0	L
Bk1	23-36	8.0	0.59	13	0	SIL
Bkq2	35-64	8.1	0.40	16	0	SIL
2Bkqm	64-91	8.5	0.28	11	0	L
3C	91+	8.4	0.18	3	0	FSL
Unnamed silt loam: Coarse-silty, mixed, mesic Xerollic Haplargids - h.t.-*A. wyomingensis / Stipa thurberiana*						
A	0-8	7.0	0.69	1	0	SIL
Bt1	8-13	6.9	0.64	1	0	L
Bt2	13-23	7.1	0.60	1	0	SIL
Bt3	23-43	7.1	0.50	1	0	SIL
Bk1	43-64	8.1	0.31	29	0	SI
Bk2	64-127	8.4	0.25	21	0	SIL
C1	127-216	8.3	0.16	15	0	SI
C2	216-274	8.0	0.18	16	0	SI

[1]pH determined on saturated paste.
[2]O.C. determined by $K_2Cr_2O_7$ digest and $FeSO_4$ titration.
[3]CCE determined by acid neutralization and subsequent back titration with a standardized base.

The exact effect of carbonate concentration near the surface on native plant communities is not known. However, free carbonates control phosphorus at low levels due to calcium phosphate precipitates and low available micronutrients due to high pH. *A. nova* h.t. always occurs on these carbonate rich soils in Idaho (Table 1). If soils from calcareous parent material supporting *A. nova* change to acidic parent material, *A. nova* drops out.

Pedon discontinuities and duripan formation

Discontinuities caused by change in coarse fragment content and textural contrast affect soil hydrology. These textural discontinuities occur where strongly developed continuous duripans form. It is known that porosity difference, caused by abrupt textural changes, induced a build-up of soil water. With the relatively low hydrologic build up, water accumulates here and, upon evaporation carbonates and silica precipitate and accumulate the cement for duripan development. This hypothesis is verified by Hauxwell's (1977) soil moisture studies of the cryic soils.

These discontinuities and duripan formations affect the distribution of the h.t. When they are at shallow depths, the soils support *A. arbuscula* and *A. nova*. At greater depths or when entirely absent, the soils support *A. wyomingenesis, A. tripartita,* and *A. vaseyana*.

Periglacial influence on soil properties

The highly variable coarse fragment content, and to a lesser degree, the silts in cryic soils are attributed to periglacial activity during the Pinedale glacial period. Periglacially mixed materials are dominantly gravelly and very gravelly loams. Substratum textures are commonly a cobbly very gravelly loamy sand or sand. As noted previously, high percentages of gravels and cobbles appear at shallow depths in soils supporting *A. arbuscula* and *A. nova*.

The variation in soil properties caused by the formation of patterned ground or mounds and intermounds causes contrasting complex patterns of vegetation. In mound-intermound sequences these h.t. are found: *Atriplex conferifolia / A. nova, A. wyomingensis / A. nova, A. wyomingenesis / A. arbuscula* (Hauxwell, 1977). The occurrence of these complex patterns of h.t. reflects the same soil properties found in the broader, more extensive vegetation patterns associated with specific soils.

Summary and Conclusions

In Idaho, Aridisols with cryic soil temperature regimes are carbonate rich and show little evidence of clay translocation. These soils have developed on fan terraces and inset fan terraces formed from glacial outwash from limestone. The terraces range in age from Bull Lake (140,000 to 70,000 yrs) to Pinedale (70,000 to 10,000 yrs) and younger.

There are several horizon sequences represented by several combinations of Bk, Bkq, Bqk, Bqkm, and C horizons. Contrasting pedon properties, such as variable degree of duripan cementation, depth to cementation, thickness of solum, variation in particle size distribution, and gravel content contribute to different taxa, horizon sequences, and distribution of plant communities referred to as habitat types (h.t.).

Periglacial climates have resulted in the formation of patterned ground, or mound and intermound surface relief. These periglacial processes have had a major influence in the development of the contrasting pedon properties previously mentioned. The dominant soils occurring on the alluvial fans have duripans grading to petrocalcics.

Cemented regions consist of a clast-supported framework of rounded gravels and cobbles with variable amounts of matrix material. The appearance of cemented horizons changes with age and are modified by periglacial effects. The nature of cementation in these duripans is characterized by three distinct regions of cementation: a laminar cap, bridging of framework clasts by matrix material, and coalescing pendants.

The laminar caps are laminations and striations of alternating micritic calcite with brown microgranular calcite-opaline silica material. Scanning electron microscope observations show the opaline silica occurs in web-like arrangements that bridge particles and occur as isolated units. This intimate association of calcite with opaline silica is also characteristic of mature duripans, but the cemented region is thicker with the framework clasts cemented together.

The major source for matrix material filling voids among the framework clasts is eolian dust. This is also a source for silica, as well as some carbonate, and is a source for smectite in younger pans, which later alters to sepiolite and polygorskite in older pans.

The habitat type plant community classification used here is ecologically significant and is based on climax vegetation. The h.t.s are arranged along on an elevational moisture and temperature gradient. Their distribution is also a response to variations in soil properties. The ecological distribution of sagebrush occurring in the h.t.s and corresponding soil taxa given in Fig. 2. are *A. wyomingensis, A. nova,* and *A. arbuscula* on Aridisol soils and *A. arbusula , A. tripartita*, and *A. vaseyana* are on Mollisol soils. The dates of moisture depletion to 1.5 MPa was June 22 for the h.t.s in the Aridisols, July 6 for the *A. tripartita*, and July 20 for *A. vaseyana*.

Aridisols with frigid soil temperature regimes occur at lower elevations below cryic soils and above the extensive mesic soils of the Snake River Plain. The major area of frigid soils discussed here occurs at the upper and east end of the Snake River Plain. This area is dominated by two ages of calcareous loess having its source from glacial outwash during Bull Lake and Pinedale glacial episodes. The northern edge is influenced by high carbonate alluvium from the limestone formations in the cryic soil area to the north. The loess soils have carbonate leached from the A and Bw into the Bk. These soils have no duripan development and have not been effected by periglacial influences and patterned

ground formation. The sagebrush species composing the h.t.s are the same as for cryic soils. However, *A. wyomingensis* is dominant but with different grass components on the loess soils.

The O.C. in the ochric epipedon gradually increases when progressing from mesic through the frigid and cryic Aridisols. This is not evident from relatively uniform color of the ochric epipedon; however, the thickness may increase. Both properties are probably masked by high carbonates but contribute to the xerollic subgroup classifications of these Aridisols.

The abundant carbonate accumulations from carbonate rich parent materials in the cryic soils is the most striking feature. It contributes to the formation of calcic horizons, duripans, and carbonatic family classifications. The effect of the discontinuities caused by concentration of coarse fragments appears to effect the soil's hydraulic properties and partially determines the position of duripan development.

The solum depth and variations in particle size, distribution, carbonate content, and different degrees of cementation of non-indurated horizons was caused by periglacial climates and patterned formation. In the frigid areas, not influenced by periglacial climates, the formation of a cambic horizon represents a young stage of soil development correlating to less that 70,000 years before present.

Literature Cited

Blank, R.R. 1987. Genesis of silica and calcite cemented soils in Idaho. Univ. of Idaho. Ph.D.Thesis (unpublished).

Blank, R.R. and M.A. Fosberg. In press, a. Micromorophology and classification of secondary calcium carbonate accumulations that surround or occur on the undersides of coarse fragments in Idaho (U.S.A.). Proceedings of the Internation Working Meeting on Soil Micromorphology.

Blank, R.R. and M.A. Fosberg. In press, b. Duripans of Idaho, U.S.A.: In situ alteration of eolian dust (loess) to an opal-A/X-ray amorphous phase. Geoderma.

Blank, R.R. and M.A. Fosberg. 1989. Duripan of the Owyhee Plateau region of Idaho: Micromorphological evidence for polygenesis. In Agronomy Abstracts: Annual Meeting of the Amer. Soc. Agron. 258 Abstract

Chadwick, O.A., D.M. Hendricks, and W.D. Nettleton, 1987. Silicification of Holocene soils in northern Monitor Valley. Nevada Soil Sci. Soc.Am. J. 53:158-164.

Dort, W. 1969. Multiple glaciation of southern Lemhi mountains, Idaho. Preliminary reconnaissance report. Idaho St. Univ. Tebiwa, Vol. 5:2-17.

Flach, K.W., W.D. Nettleton, L.H. Gile and J.G. Cady. 1969. Pedocementation: induration by silica, carbonates and sesquoxides in the quaternary. Soil Sci. Vol. 107:442-453.

Flach, K.W., W.D. Nettleton, and R.E. Franklin. 1974. The micromorphology of silica-cemented soil horizons in western North America. In: G.K. Rutherford (Editor), Soil Microscopy. Limestone Press, Kingston, Ontario. pp. 715-729.

Fosberg, M.A. and M. Hironaka. 1964. Soil properties affecting the distribution of big and low sagebrush communities in southern Idaho. Am. Soc. of Agron., Forage Plant Physiology and Soil - Range Relationships Spec. Pub. No. 5.

Fosberg, M.A. 1965. Characteristics and genesis of patterned ground in Wisconsin time in a chestnut soil zone of southern Idaho. Soil Sci. Vol 99:30-37.

Fosberg, M.A. and C.L. McGrath. 1988. Idaho soil moisture and temperature map. Univ. of Idaho, Coll. of Agr. (unpublished).

Funk, J.M. 1977. Climatic and tectonic effects on alluvial fan systems, Birch Creek Valley, east central Idaho. Univ. of Kansas, Ph.D.Dissertation (unpublished).

Harkness, A.L. 1983. Silica cementation in soils on the Owhyee high plateau, Idaho. Soil Survey Horizons, Vol. 24, No. 2:3-6.

Hauxwell, D.L. 1977. Sagebrush soils study Birch Creek Valley, Idaho. Univ. of Idaho, College of Forestry, Range, Wildlife. Range Mgt. Spec. Res. Report (unpublished).

Hipple, K.W. and G.P. Butler. 1983. Cryic Aridisols on outwash fans and terraces in Idaho. Soil Survey Horizons, Vol. 24, No. 3:28-33.

Hipple, K.W., G.H. Logan and M.A. Fosberg. 1987. Aridisols with cryic soil temperature regimes. Proceedings of ICOMID IV.

Hironaka, M., M.A. Fosberg and A. H. Winward. 1983. Sagebrush - grass habitat types of southern Idaho. Univ. of Idaho, Coll. For. Wildlife, Range Sci. Bull. No. 35.

Jones, L.H.P. and E.R. Segnit. 1971. Nature of opal:I, Nomenclature and constituent phases. T. Geol. Soc. Aust. 18:57-68.

Knoll, K.M. 1973. Chronology of alpine glaciation stillstands east-central Lemhi range, Idaho. Univ. of Kansas, Ph.D. Dissertation.

Krantz, W.B., K.J. Gleason and N. Craine. 1988. Patterned ground, A common physical phenomenon shapes these uncommon manifestations of natural geometry. Scientific Amer. December, p. 68-76.

Lewis, G.C. and M.A. Fosberg. 1982. Distribution and character of loess and loess soils in southeastern Idaho, in Bill Bonnichsen and R.M. Breckenridge, editors, Cenozoic Geology of Idaho: Idaho Bureau of Mines and Geology Bull. 26.

Malde, H.E. 1964. Patterned ground in the western Snake River Plain, Idaho, and its possible cold climate origin. Bull. Geol. Soc. Am. 75:191-208.

Pierce, K.L., J. D. Obrodovich and I. Friedman. 1976. Obsedian hydration dating and correlation of Bull Lake and Pinedale glaciations near West Yellowstone, Montana. Geol. Soc. Am. Bull., Vol. 87:703-710.

Pierce, K.L., M.A. Fosberg, W.E. Scott, G.C. Lewis and S.M. Colman. 1982. Loess deposits of southeastern Idaho: Age and correlation of the upper two loess units, in Bill Bonnichsen and R.M. Breckenridge, editors, Cenozoic Geology of Idaho:Idaho Bureau of Mines and Geology Bull. 26.

Price, L.W. 1972. The periglacial environment, permafrost and man. Assoc. of Am Geog. Resource Paper No. 14.

Ray, R.J., W.B. Krantz, T.N. Caine and R.D. Gunn. 1983. A model for sorted pattern-ground regularity. Jour.Glaciology, Vol. 29, No. 102:317-337.

USDA-SoilConservation Service.1973. Soil survey of Bingham County area, Idaho.

USDA-SoilConservation Service. 1975. Soil Taxonomy. Ag. Handbook No. 436.

USDA-Soil Conservation Service. 1979. Soil survey of Jefferson County, Idaho.

USDA-Soil Conservation Service. 1981.Soil survey of Bonneville County area, Idaho.

UDSA-SoilConservation Service. 1984. General soils map, Idaho. USDA, Soil Cons. Service, Boise, ID.

UDSA-SoilConservation Service. 1986. Part 607, Glossary of Landform and geologic terms. pp. 607-1 to 607-45.

Washburn, A.L. 1980. Geocryology: A survey of periglacial processes and environments. John Wiley and Sons, Inc.

Altocryic Aridisols in China

Gong Zitong and Gu Guoan[1]

Abstract

Altocryic aridisol is a suborder of Aridisols in the Chinese Soil Taxonomic Classification, which means Aridisols with a frigid or colder soil temperature regime. Three great groups, i.e., frost desert soils, cold desert soils, and cryic calcic soils, are distinguished, which are concentrated in the Qingzhang Plateau of China.

The Qingzhang Plateau, the especially high and steep Kulakunlen Mts. and Kulun Mts., is a virgin land which people yearn for and see as mysterious. Until the end of 19th century only some scholars abroad and explorers such as Sven Hedin had set foot in the region to investigate geology, geography, and biology. Investigation groups from Germany, Austria, Soviet, Holland, and Italy engaged in multi-subject survey to Permir, the Karakorum Mts., and adjacent regions during the 1920s and 1950s. Liou Shene, a Chinese botanist, made an on-the-spot observation of flora in the line of the Karakorum Mts. and the Arkesai Basin of the Kunlun Mts. in 1932. After 1949, the Comprehensive Scientific Investigation Group of the Qingzhang Plateau, Academia Sinica, carried out overall explorations in Tibet and the Hendan Mts. Furthermore, multi-subject surveys began to be undertaken in details in the Karakorum Mts. and Kulun Mts. region in 1987.

This paper discusses the distribution, formation, diagnostics, types, and keys of Altocryic aridisols in this area.

Soil Distribution

Frost desert soils are mainly located in the north Kulun Mts. and the south Beiyangtang plateau, at elevations of over 5,000 m in the west and over 4,000 m in the east. The cold desert soils display on both sides of the mountains. The east side is the Chaidamu Basin, with a height of 3,000 m, and middle and low parts of the surrounding mountains, with heights of less than 4,000 m. The west side is river valley, lakesides, and mountains (<4,500 m) on the southern Kalakunlun Mts. and on the northern Ayila and Gangdesi Mts.

Cryic calcic soils are concentrated between the southern part of the frost desert soil region and the Gangdisi Mts., with elevations of 4,300 - 5,300 m (Fig. 1), and on the north slope of the middle and western Himalayas, with elevations of 4,100 - 5,100 m. They also are found in the vertical belts of the Pamir Plateau within 3,500 - 4,300 m, on the north slopes of the Kulun Mts. and Arjin Mts. at the height of 3,300 - 4,000 and 3,800 - 4,200 m, respectively, and on the south slope of Tian Mts. at the height of 2,400 - 3,000 m.

Soil-Forming Characteristics

The Qingzhang Plateau has lifted since the end of the Pliocene, resulting in high and broad relief, with an elevation of about 4,500 m and a total area of 2.4 million square km, amounting to 1/4 of the country. Moreover, the plateau is the highest, the most complex, and the youngest in the world. Above it, are rolling mountains, glaciers, frozen soils, and large basins. Because the mountains provide a natural defense for marine air current, the climate becomes drier and is characterized by cool temperatures in summer, very cold temperatures in winter, few rains, strong wind, and long frost periods. Due to relief differentiation the temperature zones can be divided into three types: the plateau arid cold zone, the plateau arid temperate zone, and the plateau semi-arid cold zone (Table 1).

The Qingzhang Plateau was invaded by glaciers extensively in the Quaternary, which lead to weathering crusts such as infancy, clastic, saltbearing and clastic carbonate.

As to vegetation, Certoides Compacta or mated by Carexmoorcroftii appears in the cold desert zone, Ceratodes mated by Kalidium Schrenkianum in the temperate desert zone, and Stipa Purpurea, S. Subsessiliflora Var. Basiplumosa etc. in the cold grass zone.

Otherwise the soil is shown with shallow solum, coarse grain size, low clay content, and low biomas. It is infertile. Gypsum-salinization, calcification, and freezing and thawing play main roles in the soil-forming process.

Gypsum-salinization

Gypsum-salinization is an important soil-forming process of frost desert soils and cold

[1]Institute of Soil Science, Academia Sinica, Nanjing.

Table 1. The climate of Altocryic Aridisol region

soil type	elevation (m)	temperature (°C) annul ave.	rainfall >0°C accum.	wind (mm)	(m/s)
Frost desert soils	4,900	-7.7	341.2	24.0	4.5
Cold desert soils	3,173	1.1	1,947	82.0	2.1
Cryic calcic soils	4,415	0.1	1,497	166.1	---

desert soils. These soils do not leach strongly in cold or temperate arid zones. Therefore gypsum and soluble salts are left or leached slightly. The soluble salts can move down or up over the gypsum horizon, because the former is more soluble to precipitate than the latter. In general, distribution of soluble salts in the profile decreases downward in a T-shape, whereas gypsum concentrates under the vesicular crust horizon, with the highest concentration in the middle of the profile in a rhombus shape, which has thickness of 10-20 cm or even more than 30 cm and a content of more than 15%, good crystalline.

Calcification

Calcification also is an important soil-forming process of cryic calcic soils. Compared with the other two soil types, the leaching and biological cycle of these soils are stronger, resulting from more rainfall and vegetation. Salts and gypsum are mostly leached away and thus carbonate leaching and precipitation become prominent. Soft and white spots and new formations of hypha carbonate can be found in the calcic horizon. Nevertheless, the calcification is relatively weak because of the long frozen period, weak chemical weathering, and weak soil-forming process.

Freezing and thawing

It indicates all actions that water freezes and thaws alternately in soil horizons are bare rocks, which gives deep effect on frost desert soils forming.

Marginal topography, such as stone-sea, stone-belt, and stone-ring, is an outcome of freezing and thawing. Soil in such areas is shallow, more gravelly, and infertile.

Another outcome is vesicular crust. When soils are frozen in the night during the warm season, vaporous water and capillary water move and condense from the subhorizon to the surface, but in the day the surface soil is thawed. CO_2 repeatedly extrudes and escapes in a horizontal direction, leading to the formation of vesicular crust.

Diagnostic Horizon and Diagnostic Features

Soil moisture regime is a leading factor governing Aridisol formation. It therefore is taken as a main criterion in classifying soil groups (Table 2).

In addition, carbonate accumulated horizon (calcic horizon), gypsum accumulated horizon (gypsic horizon), and soluble salt accumulated horizon (salic horizon), which are deeply related to soil moisture and temperature regimes are selected as main diagnostic horizons or characteristics for classifying soil subgroups.

Table 2. Identification of altocryic aridisols

Soil group	soil moisture regime (drying degree)	soil temperature (°C) annual average	the hottest
Frost ones	>10	<0	<5
Cold ones	>10	<8	>15
Cryic calcic ones	3.5-10	<8	<15

Calcic horizon

A horizon accumulation of $CaCO_3$ (sometimes $MgCO_3$) appears in the B or C layer. In semi-arid regions with a little more rainfall and stronger biological activities, carbonates may be dissolved, leached, and precipitated into the horizon, which meets the following requirements:

(1)A thickness of 15 cm or more;

(2)If carbonates in underside horizon are less than calcic horizon, carbonates in calcic horizon should be more at least 5% than C horizon (in absolute content); or

(3)If carbonates in underside horizon are more than calcic horizon, visible secondary carbonates as concretions and powder limes in calcic horizon should more than 5% or more at least 5% than surface horizon (in absolute content).

If the soil could not fully meet the above requirements but has carbonates leaching and accumulating features, it is considered calcic.

Gypsic horizon

A gypsic horizon is rich in secondary sulfate, mostly under an arid climate, and its parent materials are rich in gypsum. Gypsum crystal is coralloid in shape in a thin section (Photo 1). It meets following requirements:

(1)A thickness of 15 cm or more;

(2)Gypsum content is at least 5% more than that of the C horizon (absolute content), or the product of gypsum content in percent and the thickness of the horizon is not less

than 150; if no gypsum is in underside layer, gypsum in the 30 cm of solum should be more than 5%, and if gypsum in underside is 0-1%, gypsum in the 30 cm of solum should be more than 6%.

If the soil could not fully meet the above requirements but has gypsum leaching and accumulating features, it is considered gypsic.

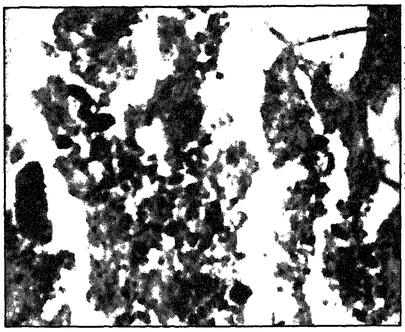


Photo 1. Gypsic frost desert soils. (K88-63), depth 6-17 cm isochromatic polarized light X 20.

Salic horizon

A soil horizon is accumulated with salts more soluble than gypsum, resulting from relics of rock weathering, precipitation or surface water migration. It appears under or in the same horizon with gypsum, and, the more arid climate, the higher position and salt content it displays. It meets flowing requirements:

(1)A thickness of 15 cm or more;

(2)Salt content is at least 2% or more but less than 50%.;

(3)The product of salt content in percent and its thickness in is 60 or more.

If the soil could not fully meet above requirements but has salts leaching and accumulating features, it is considered salic.

Freezing and thawing feature

In frost desert regions, especially near ice margins, there are disturbed forms such as stone-rings, heaving hills, lobate mud flows and takirs, etc. A vesicular crust appears in the surface and a squamose structure in the middle or low part of the cryic calcic soils.

Micromorphologically, aggregates and sorting can be seen in Photo 2 and Photo 3 respectively, and a lot of fiber clay with optical orientation can be seen in Photo 4.

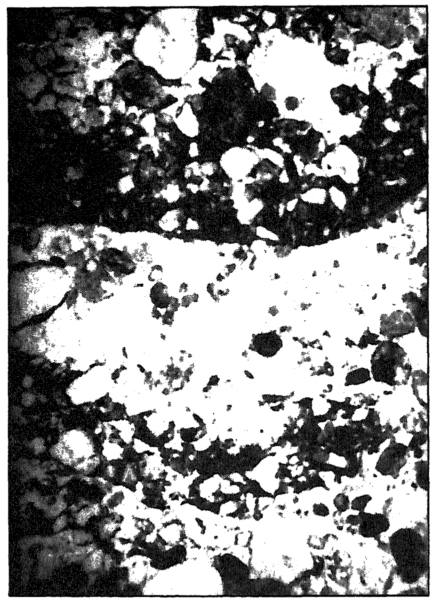
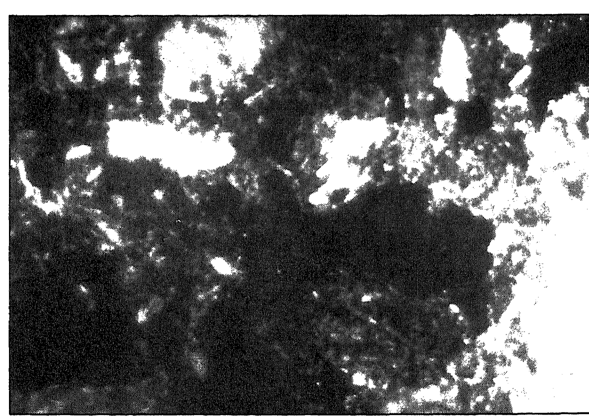


Photo 2 (above). Aggregate in cryic calcic soils, isochromatic polarized light X 20.

Photo 3 (left). Sorting in cryic calcic soils, isochromatic polarized light X 20.

Photo 4 (below). Gypsic frost desert soils, right-angled polarized light X 100.



Types

Frost desert soils

Frost desert soils are derived from clastic and salt-bearing weathering crusts, above the height of 4,000-5,000 m. The climate is cold and dry, belonging to the pergelic zone. In summer the mean air temperature is more than 0°C, and freezing occurs in the evening. Altocryic desert plants are dominant. The coverage is less than 5% or even bare. With bracteata around an ice belt, the land surface is covered by gravel or shows stone-ring and takyri of 10 cm. The basic soil-forming process is desertification, with the influence of freezing and thawing. Soil horizons are shallow, generally less than 50 cm, and abundant in gravels (more than 60%). Clay

DISTRIBUTION OF ALTOCRYIC ARIDISOLS IN CHINA

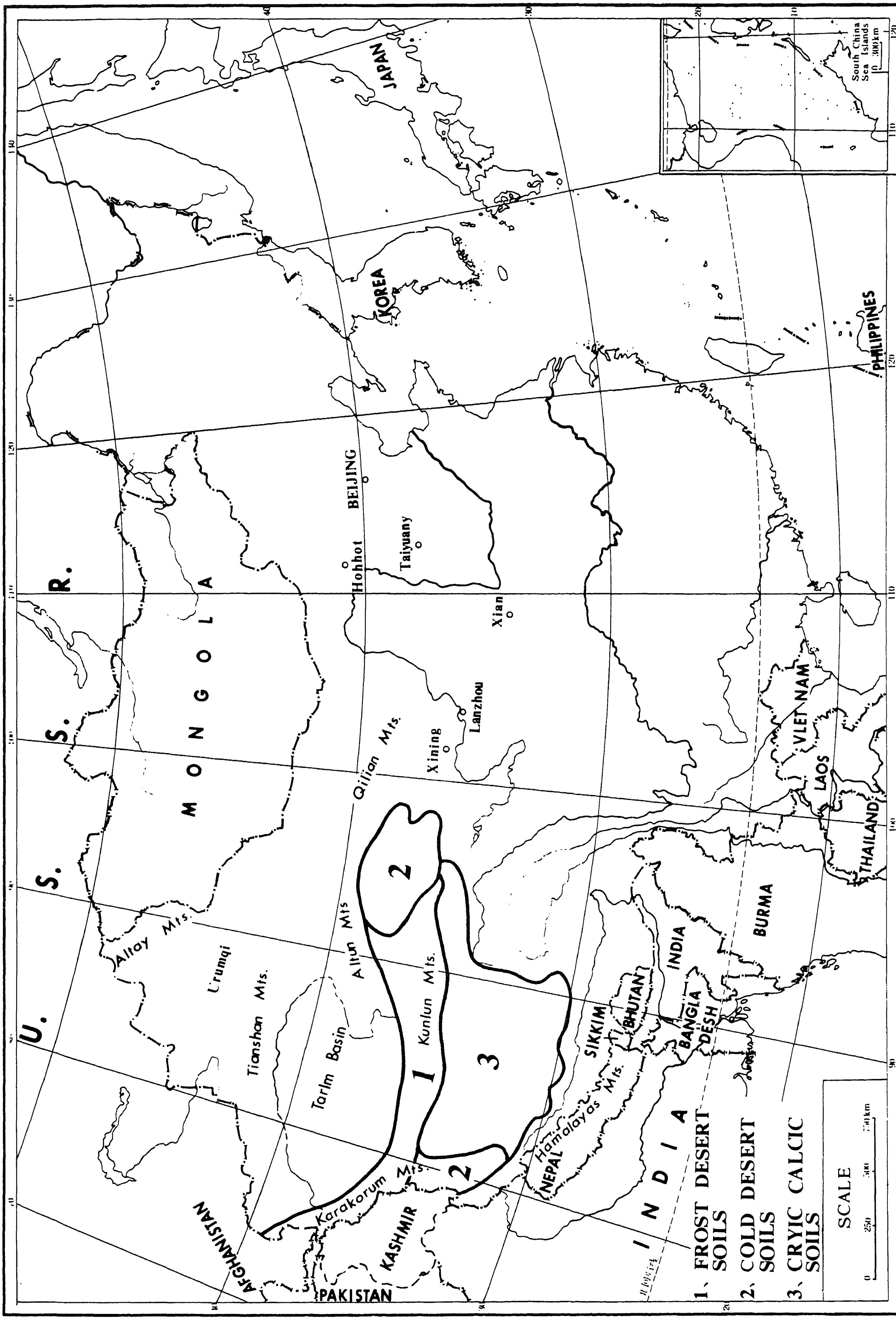


Table 3. Physical and chemical properties of frost desert soils

Soil types	depth (cm)	pH (H_2O 1:2:5)	O.M. (%)	Tot.N (%)	K_2O (%)	gypsum (%)	$CaCO_3$ (%)	salts (%)	CEC (me/100g)	Fe_2O_3(%) free	active	clay <2um
typical frost desert soil	0-4	9.0	0.41	0.024	2.36	0.51	7.74	0.019	1.79	1.69	0.20	6.1
	4-16	8.7	0.63	0.048	2.71	0.50	3.01	0.017	3.95	1.80	0.26	14.9
	16-30	8.6	0.59	0.043	2.71	0.66	1.96	0.015	3.29	2.16	0.26	10.7
	30-55	8.7	0.50	0.027	2.45	0.82	4.13	0.011	2.75	1.86	0.21	7.6
Gypsic frost desert soil	0-2	8.6	0.31	0.020	2.31	11.52	7.99	1.008	1.83	0.33	0.07	4.8
	2-6	8.6	0.42	0.025	2.55	4.80	10.24	1.152	3.22	0.45	0.12	9.1
	6-23	8.6	0.45	0.025	1.65	11.85	7.99	1.152	2.34	0.34	0.06	1.6
	23-40	8.7	0.37	0.023	2.00	8.70	11.98	0.988	2.44	0.37	0.09	1.7
	40-60	8.8	0.19	0.013	2.36	3.88	10.94	0.965	2.19	0.43	0.06	2.8
Takir Frost desert soil	0-8	8.7	1.15	0.049	1.88	7.10	25.13	0.425	3.61	0.63	0.37	19.8
	8-20	9.2	0.77	0.035	1.94	3.05	26.90	0.131	3.90	0.77	0.29	17.6
	20-32	9.2	0.45	0.028	2.15	4.87	23.00	0.070	4.01	1.00	0.75	17.6
	32-60	9.1	0.37	0.027	2.23	2.56	31.12	0.047	4.28	0.96	0.84	27.5

Table 4. Physical and chemical properties of cold desert soils

Soil types	depth (cm)	pH (H_2O1:2:5)	O.M. (%)	Tot. N (%)	K_2O (%)	gypsum (%)	$CaCO_3$ (%)	salts (%)	CEC (me/100g)	Fe_2O_3(%) free	active	clay <2um
Gypsic cold desert soil	0-1	8.5	0.22	0.019	2.17	2.25	11.28	0.487	2.37	0.48	0.09	7.5
	1-6	8.5	0.26	0.019	2.13	6.33	11.28	0.815	3.07	0.60	0.10	15.4
	6-17	8.8	0.24	0.014	1.56	17.78	9.03	1.503	1.85	0.36	0.06	4.6
	17-28	8.7	0.28	0.010	1.48	23.13	6.25	1.627	1.93	0.41	0.06	4.2
	28-50	8.8	0.24	0.010	1.83	15.73	8.68	2.114	1.71	0.52	0.09	2.4
	50-70	8.8	0.20	0.010	2.41	5.11	9.37	0.870	2.00	0.60	0.10	0.9
Salic cold desert soil	0-3	8.6	0.73	0.021	2.28	1.76	12.88	5.555		0.51	0.10	11.8
	3-11	8.8	0.49	0.016	2.25	2.64	12.01	4.697		0.47	0.10	9.7
	11-30	8.9	0.27	0.015	2.46	2.09	12.53	1.739		0.43	0.08	8.5
	30-50	8.8	0.23	0.013	2.23	1.17	10.61	1.646		0.37	0.06	4.4

content is less than 10% except for Takyri frost desert soils derived from lake-deposit. Capacity is 2-4 m.e./100g soil. The surface horizon has a vesicular crust of red brown color due to iron oxide found in the regions of especially dry and ancient, downward, gypsum accumulation horizons or where salt accumulation occurred. Clay and free and active iron contents in the depth of 2-6 cm of gypsic frost desert soils are especially high (Table 3).

O.M. accumulation is weak in frost desert soils, usually less than 0.6 percent dominated by fulvic acid. The soil is alkaline in reaction. K_2O content varies from 1.6 to 2.7% in fine earth fraction and is more than 5% in clay fraction. Clay minerals are dominated by chlorite and hydrous mica. Boron is abundant in the soil.

The region of frost desert soils is not suitable for grazing because of the low productivity of grass and far distance from pastoral areas, but it is suitable for the breeding of wildlife such as Tibetan antelope and Asiatic wild ass.

Cold desert soils

Located on a plateau of temperate desert, cold desert soils have intensive desertification and salinization, while thaw and freeze reaction is relatively weak. The soil is covered by gravel, shows white salt, and has a shallow profile generally less than 80 cm. Clay content varies depending on parent materials. CEC changes between 2-4 m.e./100 g soil. The surface horizon is characterized by a vesicular crust several cm thick. The subsurface horizon has a relatively small change of moisture and temperature regimes. Therefore prompt mineral weathering has resulted in the highest contents of clay and free and active iron oxides (Table 4). Distribution of gypsum, salt, calcium carbonate, and clay content in the profile are shown in Fig, 2.

Most of the cold desert soils are used for animal husbandry. Some have been used for agriculture when the horizons are thick and irrigation is possible.

Cryic calcic soils

Cryic calcic soils are located in cold and dry regions. Compared with froze desert soils and cold desert soils, they are characterized by (1) relatively high rainfall of 100-300 mm; (2) vegetation composed of perennial plants, having a relatively large coverage and (3) loess parent material with no stone and gravel in the surface horizon. Leaching and biological processes in these soils were strengthened. Soluble salts and most gypsum were leached out. Calcium carbonate moved down and clayification evidently occurred. Semi-decomposed O.M. residual can be seen in the surface horizon. Fulvic acid slightly dominated in the humus. CEC varies from 3 to 7 m.e./100 g soil, depending on clay and O.M. content (Table 5). Clay minerals were

dominated by hydromicas, with some chlorite.

Cryic desert soil regions are suitable for pasture but with two constraints. Pasture far from a village was under-used and pasture around a village was over-used.

Table 5. Physical and chemical properties of cryic calcic desert soils

Soil types	depth (cm)	pH (H_2O 1:2:5)	O.M. (%)	Tot. N (%)	K_2O (%)	gypsum (%)	$CaCO_3$ (%)	salts (%)	CEC (me/100g)	clay <2um
Greyic cryic desert soil	0-11	8.3	2.44	0.120	2.13	0.53	12.26	0.397	7.02	11.7
	11-23	8.6	1.50	0.082	2.18	0.63	12.94	0.028	6.19	11.7
	23-60	8.8	1.75	0.091	2.17	0.78	15.79	0.032	6.68	12.9
	60-83	8.5	1.49	0.075	2.02	3.57	13.94	0.114	5.51	13.6
	83-110	8.2	1.78	0.124	2.09	3.90	14.45	0.313	6.44	13.4
Salic cold desert soil	0-11	9.0	1.21	0.070	1.48	0.27	11.14	0.027	3.22	6.8
	11-25	9.0	1.00	0.069	1.73	0.38	15.49	0.034	5.95	18.2
	25-48	9.2	0.31	0.027	1.61	0.25	15.31	0.026	3.41	15.6
	48-70	9.0	0.19	0.021	1.50	0.53	13.57	0.025	2.73	12.7

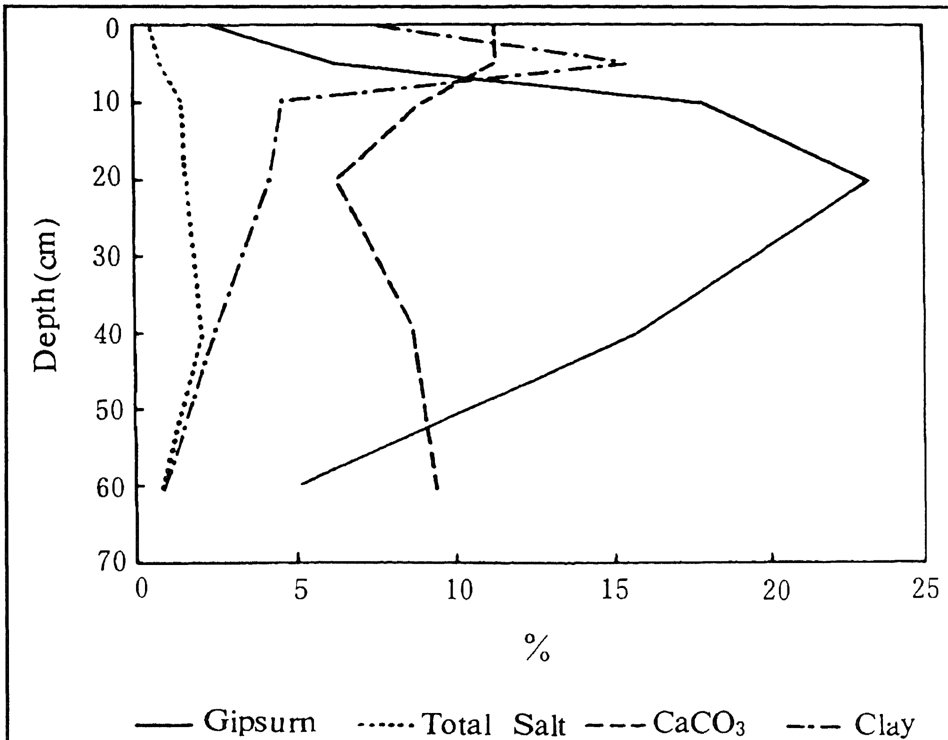


Fig. 2. Gipusm distribution in Gypsic frost desert soils

Key to Altocryic Aridisols

Key to suborder

Aridisols that have a frigid or more cold soil temperature regime

Altocryic Aridisol

Key to groups and subgroups

Altocryic Aridisols that have a pergelic soil temperature regime, vesicular crust, and freezing-thawing disturbed characteristics

Frost Desert Soil

Frost Desert Soils that meet following requirements:

(1)having no salic horizon, with upper boundary within 20 cm from soil surface;

(2)having no gypsic horizon or gypsum accumulation features, with upper boundary within 20 cm from soil surface;

(3)having no takir feature caused by freezing-thawing

- Typical Frost Desert Soil

Frost Desert Soils that meet above requirements except (1)

- Salic Frost Desert Soil

Frost Desert Soils that meet above requirements except (2)

- Gypsic Frost Desert Soil

Frost Desert Soils that meet above requirements except (3)

- Takir Frost Desert Soil

Other Altocryic Aridisols that have frigid sol temperature regime and vesicular crust

Cold Desert Soil

Cold Desert Soils that meet following requirements:

(1)having no salic horizon, with upper boundary within 20 cm from soil surface;

(2)having no gypsic horizon or gypsic accumulation features, with upper boundary within 20 cm from soil surface

- Typical Cold Desert Soil

Cold Desert Soils that meet above requirements except (1)

- Salic Cold Desert Soil

Cold Desert Soils that meet above requirements except (2)

- Gypsic Cold Desert Soil

Other Altocryic Aridisols that have cryic or more cold soil temperature regime and calcic horizon or calcic feature except for soils derived from non-carbonated parent material

Cryic Calcic Soil

Cold Calcic Soils that meet following requirements:

(1)having an ochric humus surface horizon that have a weighted O.M. content of less then 1% within 40 cm from soil surface;
(2)having no secondary clayification subsurface horizon and having cambic horizon;
(3)having calcic feature and no calcic horizon;
(4)having no calcareous reaction from soil surface to the bottom of B horizon

- Typic Cryic Calcic Soil

Cryic Calcic Soils that meet above requirements except (1)

- Greyic Cryic Calcic Soil

Cryic Calcic Soils that meet above requirements except (2)

- Clayic Cryic Calcic Soil

Cryic Calcic Soils that meet above requirements except (3)

- Calcic Cryic Calcic Soil

Cryic Calcic Soils that meet above requirements except (4)

Calcareous Cryic Calcic Soil

Main References

Lei Wenjin, 1989. Classification of Aridisols in China. Soils, Vol. 21:2 (in Chinese)

Gao Yixing et al., 1985. Soils of Tibet. Science Press, Beijing (in_Chinese)

Wen Zenwang, 1965. Soi Geography of Xinjiang. Science Press, Beijing (in Chinese)

Gong Zitong and Lei Wenjin, 1989. Aridisols of China. In the Proceedings of the Fourth International soil Correlation Meeting. pp. 111-119. USDA

Chao Sungchiao, 1984. The sandy deserts and the Gobi of China. In Deserts and Arid Lands, pp. 95-114. Martins Nizhoff Publishers.

Aridisols of Spain

J. Herrero[1] and J. Porta[2]

Abstract

This paper reviews the distribution and classification of Spanish Aridisols and discusses some taxonomic problems and major management practices. Little literature is available about Spanish Aridisols. Most of the information contained in this paper comes from soil maps and from the authors' unpublished data. The discussion covers the update of Soil Taxonomy, the unsuitability of Newhall's method for some Spanish Mediterranean areas, and the taxonomic status of the gypsiferous soils.

Many of the soils classified as Aridisols (S.S.S., 1975) are now distributed into several Orders because the EC_e criterion has been dropped at this hierarchic level. Recently, a new method for soil moisture regime calculation has been developed in Spain. This method improves the delineation of the aridic regions. Micromorphology provides a better knowledge of gypsiferous soils, and two kinds of gypsic horizons have been identified. Apart from marginal uses, the management of Spanish Aridisols is based on water supply and/or water saving and on taking advantage of early cropping.

The following conclusions can be made: (i) the new model of soil moisture regime calculations can be applied to the presumed aridic areas of Spain and calibrated with soil moisture measurements in the field, (ii) the inclusion of gypsiferous soils into three Orders disagrees with their distinct morphology and behavior, (iii) refinement in the gypsic horizon definition is needed to link microscopic features and field criteria, and (iv) both old and new management practices in Aridisols must be reported in detail for improvement, the adaptation of new technologies, and extension of management practices to new areas.

Introduction

In areas with a Mediterranean climate in Spain, irrigation is often the only way for sustainable agriculture. Historical notes show that small irrigation canals (acequias) were built more than 2000 years ago; and some 31000 Km^2 were under irrigation in 1984-85 (Leòn and Delgado, 1988).

Spain has great expanses subject to a semiarid climate, and early soil scientists such as Huguet del Villar (1929, 1950) and Kubiena (1953) created specific categories for Spanish soils associated with aridity. In many Mediterranean countries, soil moisture is limiting for agriculture, and the concept of Aridisol is useful. Although the management of arid soils is well known in Spain, their classification and mapping according to S.S.S. (1975, 1987) or to FAO (1974) still poses significant problems.

Physiological drought is the essential feature used in defining the Aridisols. They must lack plant available water for some reason, at least for defined periods, and have pedogenic horizons.

The initial concept of Aridisol included both water and salinity stress (S.S.S., 1960). The criterion of electrical conductivity (EC_e) was introduced by the 7th Approximation (EC_e > 1 dS/m at 25°C) and was maintained in *Soil Taxonomy* (S.S.S., 1975) (EC_e > 2 dS/m) but has been eliminated recently (S.S.S., 1987). At present, emphasis is given to the soil moisture regime, and now the aridic regime is required for all Aridisols, with the exception only of salic horizon occurrence.

Soil Moisture Regime as a Base for the Definition of Aridisols

The weakness in the estimation of soil moisture regime from data collected by meteorological observatories is apparent when the results of calculations are compared with the kinds and amounts of natural vegetation produced in major soil areas (Guthrie, 1985). The same disagreements were observed when applying Newhall's model to predict the soil moisture regime in some Spanish semiarid regions (Làzaro et al., 1978; Porta et al., 1983; Alberto et al., 1984; Jarauta, 1989). Spanish areas having a typical Mediterranean climate and vegetation

[1]Dept. of Soils & Irrigation, SIA-DGA, P.O. Box 727, 50080 Zaragoza (Spain). Corresponding author.

[2]Dept. of Meterorology & Soil Sci., ETSI Agrònomos, Rovira Roure 177, 25006 Lèrida (Spain).

are included in the ustic soil moisture regime after Newhall's method (Tavernier and Wambeke, 1976).

The model of moisture accretion and depletion developed for soils in the Great Prairies (S.S.S., 1975; Newhall, 1976) uses mean monthly precipitation and temperature to calculate the soil moisture regime, assuming a general available water capacity of 200 mm. The need for dichotomy in the diagnosis criteria and the consequent changes in the calculations were pointed up in Spain (Elìas and Ibànez, 1979; Gascò and Ibànez, 1979; Ibànez and Gascò, 1983). Very few field measurements of soil moisture contents are available around the world, and the same is true for Spain.

Jarauta (1989) has investigated the soil moisture contents in some sites of northeastern Spain with an ustic soil moisture regime, using Newhall's method. These sites have Mediterranean vegetation, and their crop production is affected by soil moisture shortage during the plant growing season. A new model of moisture accretion and depletion has been proposed, after four years of field measurements of soil moisture by the gravimetric method. This model has been designed to allow for the incorporation of local climate and soil characteristics.

Jarauta's model precisely defines the soil moisture regimes and allows different soil control sections and water retention capacities as well as different rain dates during the month to be considered. This model approximates better than Newhall's model to field measurements of soil moisture. Its results allow for improved knowledge of the distribution of aridic soils in Spain, separating aridic-xeric from ustic regimes. In the Ebro basin (NE Spain), this has caused the ustic regime to be rejected reference to either by soil moisture effects on soil production or four years' measurement of soil moisture content in the soil moisture control section.

The soils having argillic or natric epipedon are excluded from Aridisols if their epipedon is both massive and hard or very hard when dry (S.S.S., 1975, 1987). This is a useful criterion because the data about soil moisture regimes are often unavailable.

Areas in Spain with an Aridic Soil Moisture Regime

The application of Jarauta's method will allow a better understanding of the true extent of the aridic regime in Spain. It can be hoped that this model will be applied to more observatories and to sites where soil moisture data will be gathered. So far, the map by Làzaro et al. (1978) with some modifications can be used for Peninsular Spain.

Table 1. Diagnostic horizons in the Aridisols of Spain.

	Horizon	Frequency
Epipedons	Ochric	+ + + + +
Endopedons	Calcic	+ + + +
	Cambic	+ + +
	Petrocalcic	+ + +
	Gypsic and "Hypergypsic"	+ +
	Argillic	+ +
	Salic	+
	Natric	+

Because of the lack of field measurements, several authors have proposed pragmatic criteria in order to attempt to define the extent of the aridic soil moisture regime. Neither the selected site characteristic s nor their proposed values are agreed on by the different authors. The most common criterion is elevation (Dìaz, 1987; Iñiguez et al., 1988; Pèrez et al., 1987a) or elevation plus slope orientation (Alìas et al., 1987a,b, 1988; Torre and Alìas, 1987). In other cases the distance from the Mediterranean Sea or longitude (Alberto et al., 1984; Pèrez et al., 1987b) also are included. In the future, satellite data (Milford, 1987) or hand-held sensors may furnish valuable data about the soil moisture content of bare or range soils.

The Aridisols of Spain are distributed into three main regions, (i) the Ebro Valley in northeastern Spain, (ii) the southeastern region of Spain, and (iii) small areas in the Canary Islands. The first two regions are the most arid in Western Europe. Their vegetation is quite specific and contains species whose nearest localities are in the Eastern Mediterranean or in North Africa.

Taxa of Aridisols in Spain

The recent dropping of the electrical conductivity criterion (EC_e > 2 dS/m at 25°C) forces a review of the bibliographical references of Aridisols. Soils having an EC_e > 2 dS/m must be excluded from Aridisols if they do not fit in the aridic regime or if they do not have a salic horizon fitting the requirements for Aridisols. Most of the saline Aridisols after S.S.S. (1975) are now in several Orders, having their saline characteristics reflected at the phase level (Porta and Boixadera, 1988).

An accurate interpretation of available data about Aridisols of Spain needs a definition of the salic horizon based on EC. In practice, salinity is measured as EC, and nomograms or approxi-

mate calculations are used for conversion to salt percentage for classification purposes. The updating of the definition of salic horizon should state an EC threshold in the standard extract at the soil:water ratio of 1:5.

Table 1 shows the estimated frequencies of diagnostic horizons in Spanish Aridisols, based on a bibliographical search and the authors' field experience.

Silica cementations producing duripans do not occur in Spanish aridic areas. Calcium carbonate, gypsum, and more soluble salts move, producing specific categories of soils.

Salorthids are not common in salt-affected areas of Spain. Salorthids are associated with a shallow saline water table under an ascensional soil moisture regime; their vegetation is *Arthrocnemum glaucum* and *Salicornia sp.* Soils with salic horizon occur in small areas; they are mappable only at detailed scales and references to them are scattered in the literature.

Table 2 displays the taxa of Spanish Aridisols and their location. The table was prepared after a critical review of the bibliography, and most of the references come from LUCDEME soil maps.

Table 2. Great Groups of Aridisols cited in Spain.

Great Group	Region	Province	Area	Reference
Haplargids	S.E.	Almerìa	Las Negras	Aguilar & al.,1973
			Campo de Dalìas	Martìnez-Raya, 1987
			Macael	Aguilar & al.,1987
			Roquetas	Pèrez & al., 1987
			Los Nietos	authors
Natrargids	N.E.	Huesca	Fraella	authors
		Zaragoza	Bardenas	Martìnez-Beltràn, 1978
	Canary Islands	Tenerife	Tenerife	Rodrìguez-Hdez.& al., 1980
Paleargid	S.E.	Almerìa	Rodalquilar	Aguilar & al.,1973
			Fiñana	Aguilar & al.,1987
			Roquetas	Pèrez & al., 1987
			Tabernas	Pèrez & al., 1987
		Murcia	C. Cartagena	Gisbert, 1973
Calciorthids	S.E.	Alicante	Maigmò	Alìas & Torre, 1987
		Almerìa	Nìjar	Aguilar & al., 1973
				Porta & al., 1980
			Fiñana	Aguilar & al., 1987
			Macael	Aguilar & al., 1987
			Tabernas	Pèrez & al., 1987
			Guadix	Ortega & al., 1988
		Granada	Cùllar	Simòn & al., 1980
		Murcia	C.Cartagena	Gisbert, 1973
			Cehegin	Alìas & al., 1987
			Coy	Alìas & al., 1987
			Lorca	Alìas & al., 1988
			Puerto-Lumbreras	Alìas & al, 1988
	Canary Islands	Tenerife	Tenerife	Escobar & al., 1973
				Fdez-Caldas & al., 1978
Camborthids	N.E.	Navarra	Bardenas	Arricibita, 1987
				Iñiguez & al., 1988
	S.E.	Almerìa	Nìjar	Aguilar & al., 1973
				Porta & al., 1980
			Hùercal-Overa	Alonso, 1983
			Campo de Dalìas	Martìnez-Raya, 1987
			Roquetas	Pèrez & al., 1987
		Murcia	C.Cartagena	Gisbert, 1973
			Lorca	Alìas & al., 1988
	Canary Islands	Tenerife	Tenerife	Escobar & al., 1973
Gypsiorthids	N.E.	Navarra	Lodosa	Arricibita, 1987
			Bardenas	Iñiguez & al., 1988
		Teruel	Hìjar	Porta, 1986
	S.E.	Alicante	Maigmò	Alìas & al., 1987
		Almerìa	Tabernas	Pèrez & al., 1987
		Murcia	—	Sànchez & al.,1982
			Puerto Lumbreras	Alìas & al., 1988
			Lorca	Alìas & al., 1988
	Canary Islands	Tenerife	Gomera	Jimènez & al., 1988
Paleorthids	N.E.	Huesca	Lanaja	authors
		Lèrida	Suñer	Porta & al., 1983
	S.E.	Almerìa	Nìjar	Porta & al., 1980
			Campo de	Martìnez-
		Dalìas	Raya, 1987	
		Murcia	Cehegin	Alìas & al., 1987
			Coy	Alìas & al., 1987
			Lorca	Alìas & al., 1988
			Puerto Lumbreras	Alìas & al., 1988
	Canary Islands	Tenerife	Tenerife	Escobar & al., 1973
Salorthids	N.E.	Zaragoza	Bujaraloz	Herrero, 1982
	S.E.	Almerìa	Vega de Pulpì	Alonso, 1983
			Campo de Dalìas	Martìnez-Raya, 1987
			Roquetas	Pèrez & al., 1987
		Granada	Cùllar	Simòn & al., 1980
		Murcia	—	Sànchez & al.,1982
	Balears	Majorca	Alcudia	Porta & al., 1987
	Central	C. Real	La Mancha	Porta, 1975
		Toledo	Ocaña	Gumuzzio & al., 1984
	South	Huelva	Doñana	Ayerbe & al., 1978
		Càdiz	Puerto de Santa Marìa	Gòmez & al., 1982

Gypsiferous Soils: An "Erratic" Type of Soil

In Spain, gypsiferous materials outcrop only in the east, having an extent of 35487 km^2 (Macau and Riba, 1965). Many of these outcrops are in the aridic regions. The soils developed from gyprock and other gypseous soils are well distinguished and easily separated from saline soils, both by farmers and by early soil scientists (Huguet del Villar, 1929). The *Soil Taxonomy* approach allows reflection of the genetic and management specifities of gypsiferous soils.

Gypsiferous soils in former *Soil Taxonomy* approximations

The soils enriched with "calcium sulfate" were considered in the *Soil Survey Manual* (S.S.S., 1951) and in the 5th Approximation of *Soil Taxonomy* (Cline, 1979).

The gypsic horizon was defined in the 7th Approximation (S.S.S., 1960), but soils enriched with gypsum were included in the Aridisols, together with soils having a calcic horizon, as Orthic Calcorthids. This situation was unsatisfactory, both from a genetic point of view and for land evaluation from large scale maps.

The Great Group of Gypsiorthids was developed in the Galley Proofs of *Soil Taxonomy* (1970-1973). The new Great Group was introduced in the order of Aridisols (S.S.S.,1975). A water solution saturated with gypsum (2.6 g/l at 25°C) shows an EC > 2dS/m. Accordingly, all soils having a gypsic horizon (except those with a mollic epipedon) were included in the Gypsiorthids.

The present status of gypsiferous soils

In the last revision of *Soil Taxonomy* (S.S.S., 1987), the soils with a gypsic horizon can belong to three different Orders: those with a xeric moisture regime and a mollic epipedon are Calcixerolls; those without a mollic epipedon are Xerochrepts; and those with an aridic moisture regime are Gypsiorthids. This situation increases the heterogeneity of the Great Group of the Xerochrepts, because a new Subgroup must be created: Gypsic Xerochrepts. Both the current definition of Gypsiorthids and that proposed for Gypsids (ICOMID, 1989) are based on their moisture regime. This results in problems in the location of these soils because soil moisture calculations must be made from scarce field measurements of soil moisture.

Soil moisture regime criteria may not be good enough to discriminate soil behavior and land use when a well developed, massive gypsic horizon is present. Most of these horizons may be classified as hypergypsic (ICOMID, 1989) and under xeric or aridic regimes are impenetrable for roots in the dry season (Porta et al., 1977). The placement of these gypsiferous soils into different Orders (Aridisols and Inceptisols) is not fully satisfactory.

The need for refinement in the gypsic horizon definition

The process of gypsum accumulation in the Aridisols can lead to highly gypsum-rich horizons. Field and micromorphological data about gypsic and petrogypsic horizons around the world has been reviewed by Herrero (1990). Tavernier et al. (1981) used the concept of hypergypsic, and Witty (1985) discussed this, pointing out that some relevant features such as subsidence or erosion that are currently associated with the gypsic horizon can be explained by the gypseous substratum and identified as a phase. It can be concluded that there is a need for refinement of the gypsic horizon definition, and worldwide research is necessary in order to give it a broad scope.

In the xeric and aridic soils studied in Spain (Porta and Herrero, 1990), the gypsic horizons in Gypsic Xerochrepts and in Gypsiorthids commonly have similar micromorphological characters. Moreover, the low water retention capability of gypsum enhances the arid conditions of these soils.

The gyprock often produces a mass of microcrystalline gypsum that can be richer in gypsum than the parent gyprock. So, a gypsum enrichment process can be accepted, although the causes of the formation of this kind of gypsum crystals remain unknown. This material either stands on the parental gyprock or moves along the slope as mud-flow. This weathering product with an upper epipedon may be identified as a gypsic horizon from the morphology and fits the definition of hypergypsic horizon.

Terms affecting the composition and quantification criteria are misused in some cases. Gypsum and calcium sulfate are misused as synonymous, and even $CaSO_4$ (anhydrite) is employed instead of $CaSO_4 \cdot 2H_2O$ (gypsum).

Land Use of Aridisols in Spain

Non irrigated lands

The Aridisols are an important soil resource in Spain, but their moisture is usually too low to support rainfed agriculture. In dry farming, only winter cereals are possible, barley and wheat being the most common. Short duration cereals are often preferred because of drought at the end of each cycle. In some aridic areas in transition with xeric, crops of almond and olive trees are possible. In other areas, dry farming must be alternated with range for grazing sheep, or even only range may be possible. The marginal harvesting of plants such as *Lygeum spartum* for fiber and other plants for soap has been abandoned. Recreation and wildlife could be other alternative uses for these lands.

Irrigated lands

Flooding is the traditional irrigation system in Spain, as in most places in the world. In the new irrigation districts of Spain, sprinkler irrigation is commonly used for extensive crops, and trickle irrigation for fruit trees and vegetable crops.

In the northeast aridic region (Ebro basin), water of good quality was abundant, and rice and drainage were used for salt removal. Notwithstanding, salt-affected Aridisols occur in some irrigated districts of the Ebro basin (Martìnez-Beltràn, 1978; Herrero and Aragüès, 1988). Severe problems remain in soils whose high silt content, sodicity and adverse micromorphological features make pipe drainage difficult (Rodrìguez et al., 1989).

In the southeast aridic region, scattered plots began to be irrigated from small earth reservoirs built for the prevention of flash floods due to storms (Leòn et al., 1987). The increase in irrigated surface has led to some water shortage and quality problems.

Special management practices

After the end of the 19th century, a special soil management system for vegetable production was developed in the southeast of Spain and in the Canary Islands. Shortage of irrigation water and/or low water quality requires the use of this system, which is called "enarenado" (sand mulching). Many effects of "enarenado" have been cited (Martìnez-Raya, 1987): (i) reduction of water loss by evaporation, (ii) atmospheric water condensation, (iii) soil temperature regulation, (iv) advancement of harvest date, and (v) saving in fertilizer. Depending on local soil and climate circumstances or on market considerations, each of these effects can be decisive for the profitability of a crop.

In two decades, a great surface area has been developed with polyethylene covered crops in the aridic zones of southeastern Spain (Leòn and Delgado, 1988). This technique is often combined with "enarenado," and high profitability is being obtained with extra-early horticultural crops.

Conclusions

The broad distribution of Aridisols in Spain is well known. A more accurate discrimination between aridic and xeric in Mediterranean conditions has been attained with Jarauta's method of soil moisture regime calculation. The method was tested in the Ebro basin and could be useful in other areas using soil moisture measurements for control.

Some gypsiferous soils, e.g., in central Spain, that were classified as Aridisols, belong now to Gypsic Xerochrepts. This approach underlines the differences in crop production, but field and micromorphological studies show a convergence in the morphology, behavior, and management of gypsiferous soils in Spain, in spite of their classification in different Orders.

Advances in knowledge of the soil moisture regime of aridic soils must contribute to the planning of soil and water management in wide areas of Spain and to the transfer of agricultural technology.

Literature Cited

Aguilar, J. & al. 1987. Mapa de Suelos 1:100.000. Fiñana. LUCDEME. M.A.P.A. Madrid.

Aguilar, J. & al. 1987. Mapa de Suelos 1:100.000. Macael. LUCDEME. M.A.P.A. Madrid.

Aguilar, J., J. Fernàndez, A. Garcìa, and J. Ìñiguez. 1973. Estudio edàfico de la zona de Cabo de Gata. II. Estudio de Mollisuelos y Aridisuelos. An. Edaf. y Agrobiol. 32:299-340.

Alberto, F., J.L. Arru è, and J. Machìn. 1984. El clima de los suelos de la Cuenca del Ebro. I. Reg ìmenes de Humedad. Anales Estaciòn Experimental de Aula Dei 17(1-2):7- 20.

Alìas, L.J. & al. 1987a. Mapa de Suelos 1:100.000. Cehegin. LUCDEME. M.A.P.A. Madrid.

Alìas, L.J. & al. 1987b. Mapa de Suelos 1:100.000. Coy. LUCDEME. M.A.P.A. Madrid.

Alìas, L.J. & al. 1988a. Mapa de Suelos 1:100.000. Lorca. LUCDEME. M.A.P.A. Madrid.

Alìas, L.J. & al. 1988b Mapa de Suelos 1:100.000. Puerto Lumbreras. LUCDEME. M.A.P.A. Madrid.

Alìas, L.J., and A. de la Torre. 1987. Suelos de la Sierra del Maigmò (Alicante), II: Descripciòn de perfiles. Datos analìticos, clasificaciòn y distribuciòn. An. Edaf. Agrobiol. 46:909-930.

Alonso, A. 1983. Estudio de los suelos de "El Saltador" (Huèrcal-Overa) y "La Vega" (Pulpì), en Almerìa. Doc. Thesis, Univ. Madrid.

Arricibita, F. J. 1987. Tipologìa de suelos salinos en Navarra. Doc. Thesis. Univ. Navarra.

Ayerbe, M. & al. 1978. Guìa de las excursiones cientìficas. VII Reuniòn Nacional de Suelos. Sevilla. CSIC-SECS.

Cline, M.G. 1979. Soil Classification in the United States. Agron. mimeo N° 79-12. Dept. of Agron. Ithaca. 207 p.

Dìaz, J.L. 1987. Estudio edfico y evaluaciòn forestal en la Sierra de los Filabres. I.A.R.A., Sevilla.

Elìas, F., and V. Ibàñez. 1979. Comparaciòn de dos modelos matemàticos para estimar el règimen de humedad del suelo. An. INIA, Ser. General, 6:49-76.

Escobar, F., and C. Roquero. 1973. Estudios agroedafològicos de las àreas con aptitud para producir tabaco en la isla de Tenerife. Min. de Agricultura, Madrid.

FAO. 1974. Soil map of the world 1:5.000.000. Vol. I. Legend. UNESCO. Paris. 59 p.

Fernàndez-Caldas, E., and C.M. Rodrìguez-Hernàndez. 1978. Suelos formados sobre materiales volcànicos (Islas Canarias). Aridisoles: Natrargids, Calciorthids. An. Edaf. Agrobiol. 37:717-731.

Gascò, J.M., and V. Ibàñez. 1979. Criterios para la estimaciòn del règimen de humedad de los suelos. An. INIA, Ser. General, 6:61-76.

Gisbert, J. 1973. Estudio edafològico del Campo de Cartagena. Desarrollo del Orden de los Aridisoles. Doc. Thesis. Univ. of Valencia.

Gòmez, V., J. Pèrez, and R. Roquero. 1982. The soils and water table properties of polders area Castillo de Do ña Blanca, Puerto de Santa Marìa (Càdiz, Spain). Internat. Polders Symp., The Netherlands.
Gumuzzio, J., J.B. Alvarez, A. Gutièrrez, and J. Guijarro. 1984. Aproximacin a la capacidad de uso de los suelos de Ocaña (Toledo). An. Edaf. y Agrob. 43:93-111.
Guthrie, R.L. 1985. Refinement of taxa based in soil moisture regimes. Proceed. 5th Int. Soil Class. Workshop: 37-38.
Herrero, J. 1982. Salinidad del suelo en salobrares de Monegros y Somontano oscense 'como condicionante de la vegetaciòn. Inst. Fernando el Catòlico. Zaragoza.
Herrero, J. 1990. Morfolog ìa y gènesis de suelos sobre yesos. INIA, Madrid. (in press).
Herrero, J., and R. Aragüès. 1988. Suelos afectados por la salinidad en Aragòn. Surcos de Aragòn 9:5-8.
Huguet del Villar, E. 1929. Geobot ànica. Ed. Labor, Barcelona.
Huguet del Villar, E. 1950. Geoedafolog ìa. (Edition of 1983 by the Univ. of Barcelona).
Ibàñez, V., and J.M. Gasò. 1983. Modificaciones al modelo matem àtico propuesto por Newhall, F. para la estimaciòn del règimen de humedad del Suelo. Comunicaciones INIA. Serie Recursos Naturales. N. 18.
ICOMID. 1989. Aridisols, version 6.0. Internat. Committee on Aridisols.
Iñiguez, J., I. Sànchez-Carpintero, R. Val, F. J. Arricibita, M. Vidal, M. Garjòn, and G. Vitoria. 1988. Mapa de Suelos de Navarra, Hoja 207 y 245. Dep. Edafologìa. Univ. Navarra.
Jarauta, E. 1989. Modelos matem àticos del r ègimen de humedad de los suelos. Aplicaciòn al àrea meridional de Lleida. Doc. Thesis. Univ. Politècnica de Cataluña, Barcelona.
Jimènez Mendoza, C., A Rodrìguez and M.L. Tejedor. 1988. Suelos de la isla de Gomera (Canarias). II. An. Edaf. Agrobiol. 47:1159-1169.
Kubiena, W.L. 1953. The soils of Europe. Thomas Murby & Co. London.
Làzaro, F., F. Elìas, and M. Nieves. 1978. Regìmenes de humedad de los suelos de la España peninsular. Monografìas INIA, 20. Madrid.
Leòn, A., F. del Amor, and A. Torrecillas. 1987. El riego en la regiòn de Murcia. CSIC, Murcia.
Leòn, A. de, and L.F. Delgado. 1988. Memoria del Mapa de cultivos y aprovechamientos de España. M.A.P.A. Madrid.
Macau, F., and O. Riba. 1965. Situaciòn, caracterìsticas y extensiòn de los terrenos yesìferos en España. Serv. Geolgico de O.P., Madrid.
Martìnez-Beltràn, J. 1978. Drainage and reclamation of salt-affected soils in the Bardenas Area,Spain. ILRI Wageningen.
Martìnez-Raya, A. 1987. Comportamiento del riego bajo enarenado en invernadero. Doc. Thesis. Univ. Madrid.
Milford, J.R. 1987. Problems of deducing the soil water balance in dryland regions from Meteosat data. Soil Use and Manage. 3(2):51-57.
Newhall, F. 1976. Calculation of soil moisture regimes from the climatic record. USDA-SCS.
Ortega, E. & al. 1988. Mapa de suelos 1:100.000. Guadix. LUCDEME. M.A.P.A. Madrid.
Pèrez, A. & al. 1987a. Mapa de Suelos 1:100.000. Roquetas de Mar. LUCDEME. M.A.P.A. Madrid.
Pèrez, A. & al. 1987b. Mapa de Suelos 1:100.000. Tabernas. LUCDEME. M.A.P.A. Madrid.
Porta, J. 1975. Redistribuciones iònicas en suelos salinos: influencia sobre la vegeteciò n halòfila y recuperaci òn de suelos con horizonte gypsic ò y otros suelos halmorfos del rìo Gigüela. Doc. Thesis. Univ. of Madrid.
Porta, J. 1986. Edafogènesis en suelos yesìferos en medio semiàrido. Unpubl., E.T.S.I.A. Lèrida.
Porta, J. & al. 1983. Los suelos de Catalu ña: 1. Area meridional de Lèrida. Generalitat de Catalunya. Barcelona.
Porta, J. & al. 1987. Introducci ò al coneixement dels s òls. AEAC, Barcelona.
Porta, J., and J. Boixadera. 1988. Suelos y salinidad en el valle del Ebro. In I. Romagosa & al. (eds.) The basis of Crop Production. AEAC, Barcelona: 177-201.
Porta, J., and J. Herrero. 1990. Micromorphology and genesis of soils enriched with gypsum. Developments in Soil Science 19:321-339. Elsevier.
Porta J., M. Lòpez-Acevedo, and R. Darès. 1980. Los suelos del Campo de Nìjar. C.R.P. Almerìa.
Porta, J., M. Lòpez-Acevedo, and C. Roquero. 1977. Morfometrìa y clasificaciòn de algunos Gypsiorthids en España. Anales INIA, 5:85-111.
Rodrìguez, J., J. Herrero, and J. Porta. 1989. Micromorphological assessment of soil siltation risk indexes in a saline-sodic soil in Monegros irrigation district (Spain). Proc. Work. Meeting on Soil Micromorphology. San Antonio, TX, 1988.
Rodrìguez-Hernàndez, C.M., C. Rodrìguez-Pascual, and E. Fernàndez-Caldas. 1980. Aridisoles formados sobre materiales volc ànicos (Islas Canarias). Natrargids. An. Edaf. Agrobiol. 39:1415-1441.
Sànchez, J.A., and F. Artès. 1982. Gènesis, clasificaciòn y cartografìa de los suelos de la regiòn de Murcia. Caja Ahorros Provincial. 72 p.
Simòn, M., J. Aguilar, and C. Sierra. 1980. Los suelos halomorfos de la provincia de Grana da. II.- An. Edaf. Agrobiol. 39:101-120. Soil Survey Staff. 1951. Soil Survey Manual. USDA Handbook no. 18.
Soil Survey Staff. 1960. Soil classification, a comprehensive system. Soil Conserv. Serv. U.S. Dep. Agr. U.S. Govt. Print. Off. Washington DC.
Soil Survey Staff. 1975. Soil Taxonomy. USDA-SCS. Agric. Handb. 436.
Soil Survey Staff. 1987. Keys to Soil Taxonomy (3rd printing). SMSS technical monograph n° 6. Ithaca, N.Y.
Tavernier, R., A. Osman, and M. Ilaiwi. 1981. Soil Taxonomy and the soil map of Syria and Lebanon. Proc. 3rd Int. Soil Class. Workshop. ACSAD, Damascus: 83-93.
Tavernier, R., and A. van Wambeke. 1976. Determinaciòn del règimen hìdrico de los suelos de España segùn el modelo matem àtico de Newhall. Agrochimica 20(4-5):406-412.
Torre, A. de la, and L.J. Alìas. 1987. aracterìsticas bioclim àticas de interès edafol ògico de la Sierra del Maigmò (Alicante, S.E. España). Anales de Biologìa 12:29-41. Univ. Murcia.
Witty, J.E. 1985. Diagnostic horizons of carbonate and gypsum accumulations. Proc. 5th Int. Soil Class. Workshop: 269-273.

Aridisols of New Zealand

A.E. Hewitt and F.G. Beecroft[1]

Abstract

The Aridisols of New Zealand (NZ) correspond closely with the order of Semiarid Soils established in the new NZ Soil Classification. Camborthids and Haplargids predominate, and the suborder and great group limits accord well with classes established as important soil series. They occur in inland grabens in the southern South Island, where mean annual precipitation ranges from 350 to 500 mm and the temperature regime is mesic.

Argillic horizons carry many accessory characteristics but are difficult to identify in younger soils. Cambic horizons are also difficult to identify in massive, unstratified, non-calcareous loess. The distinction between Ustollic and Xerollic subgroups depends upon the ambiguous phrase "bordering on."

Permeability is an important soil quality for irrigation interpretations. Diagnostic slowly permeable and rapidly permeable layers have proved useful for defining soil classes in NZ.

NZ Aridisols are weakly weathered and have a low CEC; the resulting low buffering capacity makes them particularly sensitive to management. While this is true for NZ Aridisols, it is apparently not true for Aridisols in general. This distinction within the order could be made more explicit in *Soil Taxonomy*.

Introduction

Location and Significance

Aridisols occur in the South Island of New Zealand in the basin and range province of Central Otago (Figure 1). They are restricted to the basins and lower hills, where they are sheltered by fault block ranges from the rain-bearing westerly and southerly winds. Marked precipitation and temperature gradients occur with altitude, and Aridisols pass into Ustochrepts at the margins of the basins, which in turn pass to Dystrochrepts and Cryochrepts up the mountain slopes.

The Aridisols correspond almost exactly with the brown-grey earths of the NZ Genetic Soil Classification (Taylor and Pohlen, 1962) and the Semiarid Soils of the new NZ Soil Classification (Hewitt, 1989).

Although they cover only 2270 km^2, Aridisols are an important soil resource in the South Island. Large snow-fed rivers flow from the Southern Alps, cross the basins, and provide a source of high quality water for irrigation. Where water is unavailable, the traditional dryland extensive grazing remains. The use of irrigated land is diversifying from irrigated pasture to pip- and stone-fruit orcharding and horticulture. The climate provides the fruit industry with late-ripening, off-season products for export to northern hemisphere markets.

[1]DSIR Land Resources, Private Bag, Dunedin, New Zealand

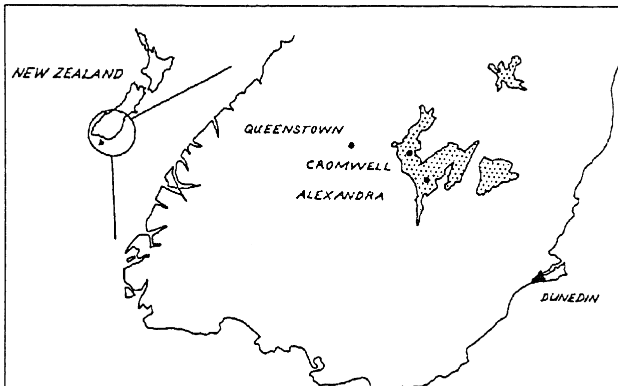


Figure 1. Location of Aridisols in the South Island of New Zealand.

Climate

Precipitation (mainly as rainfall) ranges from 330 to about 500 mm/year. It is evenly distributed throughout the year, although a slight maximum occurs in the summer. Potential evapotranspiration rates exceed 680 mm/year. The daily rate of potential evapotranspiration in summer typically ranges from 3 to 6 mm but may range from 6 to 10 mm during periods of desiccating north-west winds.

All NZ Aridisols have a mesic soil temperature regime. The mean annual soil temperature (at 30 cm depth) at Alexandra is 11.6°C, ranging from a mean monthly 3.0°C in July to 19.3°C in January.

The Newhall water-balance model was applied to data from two climate stations, Alexan-

dra and Cromwell (Watt, 1977). These gave similar results and comply with the requirements of the aridic soil moisture regime because: (1) there was a 75% probability that the moisture control section (MCS) is dry in all parts more than half the time that the soil temperature at 50 cm is above 5°C, and (2) there was only a 15% probability that the MCS is moist in some or all parts for as long as 90 consecutive days when the soil temperature at 50 cm is above 8°C.

Geology

The basement rocks are acid quartzo-feldspathic schist and weakly metamorphosed sandstone (greywacke) (Turnbull, 1987) containing no primary calcareous minerals.

In the late Cretaceous, a deeply weathered peneplain was formed on the basement rocks. Warping in the Tertiary initiated terrestrial sedimentation of coal measures, sands, silts, and clays. Block faulting in the late Pliocene and Pleistocene formed the present basin and range topography. The Tertiary sediments remain in parts of the basins but are mostly overlain by Pleistocene glacial outwash deposits. Uplands were strongly modified by periglacial action, producing tors, shaved bedrock surfaces, and solifluxion debris. Localized areas of deeply weathered schist occur on the down- throw sides of faults as remnants of profiles weathered in the late Cretaceous.

Soils and Land Units

The Aridisols occur in 4 major land units. Within these units, Aridisol suborders and great groups accord well with classes established as NZ soil series.

Terraces and Dissected Terrace Land

Terraces of rounded gravelly sand alluvium, many metres thick, occur adjacent to the major rivers. Six distinct levels are recognised, but only 4 have significantly different soils (Figure 2a). The terraces have been cut in Tertiary silts or clays. In lateral valleys, not directly affected by glacial outwash, the terrace alluvium is thinner. Irrigation in areas where it is less than about 1.5 m to the underlying sediments may cause waterlogging and salinization.

Soils on successively higher terraces increase in age and show greater soil development (Leamy, 1973). Argillic horizons are absent on lower terraces and increase in thickness from intermediate to high terraces (Figure 2a). Younger argillic horizons are usually brown (10YR hue) and older ones are usually reddish brown (5YR hue). On high terraces the argillic horizons are continuous, but on intermediate terraces they are patchy where there has been Holocene fluvial activity.

Early Pleistocene outwash deposits occur as finely dissected hills above the level of the high non-dissected terraces (Leamy 1972). Bedding in the gravels is tilted and reddish-brown argillic horizons are very thick (60-130 cm), but interrupted by younger gully fill deposits in which brown argillic horizons are absent or thin (20-30 cm).

Fans

The most extensive fans are of late Pleisotocene age and occur in association with the major terrace levels (Leamy and Saunders, 1967; Orbell, 1974).

Three distinct parts known as the apex, mid, and toe (McCraw 1968) form concentric zones in most fans (Figure 2b). In fans from schist or greywacke, the three parts are dominantly sandy-skeletal, fine-loamy and fine-loamy, or fine-silty, respectively. Soils of the fan toes are frequently imperfectly or poorly drained, and drainage may be aggravated by irrigation. Fans from catchments in Tertiary sediments are more frequently clayey, alkaline, and saline in the toe. The concentric pattern is masked in fans or parts of fans that have been overlain by loess. Fan soils are more strongly stratified than terrace soils and, where very stony, they are more likely to be loamy-skeletal than sandy-skeletal.

Tertiary Sediments and Schist Saprolite Land

Deeply weathered schist remnants of the former Cretaceous peneplain surface occur on down-throw sides of major faults (Figure 3a). The saprolite materials are of minor extent but are important for an understanding of the soils and landscape of Central Otago (McCraw, 1965). The kaolin component in soils from younger parent materials is thought to have been redistributed in the landscape from the weathered schist (Churchman 1978). The weathered schist is saline and would appear to be the ultimate source of the salts in the derived sediments.

Tertiary sediments outcrop either as smooth contoured hills or in terrace scarps. They are dominantly clayey but silts, sands, and gravels also occur. The sand fraction and coarse fragments are dominated by quartz, and the clays

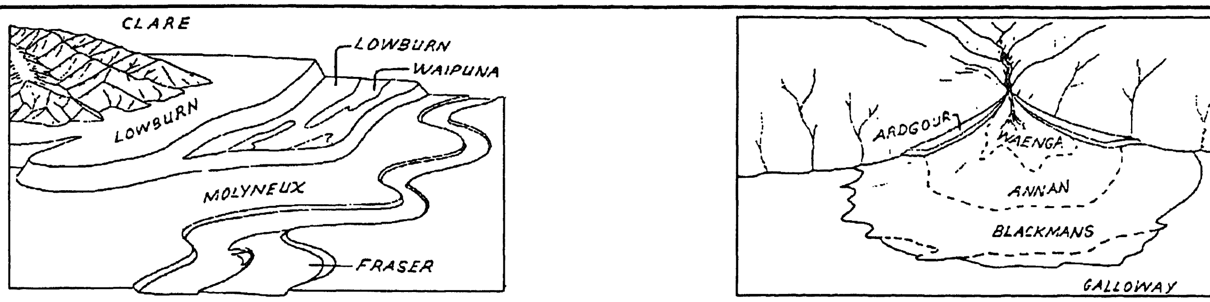


Figure 2(a) Terrace and dissected terrace land, and (b) Fan land, showing predominant soil series, their landforms, and classification

(a) **Fraser series** — Ustic Torriorthents, sandy-skeletal, m.m.*
Molyneux series — Ustollic Camborthids, sandy-skeletal, m.m.
Waipuna series — Ustollic Camborthids, sandy-skeletal, m.m.
Lowburn series — Ustollic Haplargids, loamy-skeletal, m.m.
Clare series — Ustollic Haplargids, clayey-skeletal, m.m.
*m.m. = mixed mesic

(b) **Ardgour series, Ustollic Haplargid, Loamy-skeletal, m.m.***
Waenga series, Ustollic Camborthic, Coarse-loamy, m.m.
Annan series, Ustollic Haplargid, Fine-loamy, m.m.
Blackmans series, Ustollic Haplargid, Fine-loamy, m.m.
Galloway series, Typic Haplaquept, Fine-loamy, m.m.

are predominantly kaolinite and derived from the weathered schist. A loess blanket occurs in places. The soils are slowly permeable Haplargids and in many cases are saline. Gypsum occurs in some soils where it has presumably weathered from pyrite (Lindqvist, pers comm.). The gypsiferous soils (Blakemore 1968) are the only NZ Aridisols which do not respond to sulphur fertilizers.

Basement Rocks Land

In summit, shoulder, and ridge positions, the soils are shallow Camborthids with lithic contacts. Rock outcrops are frequent with either distinctive "fretted" or "tor" landscapes (McCraw, 1965; Wood, 1969). On accumulating backslopes and footslopes, slowly permeable Haplargids occur with brown argillic horizons (Figure 3b).

Review of Taxa

The subgroups currently recognised in New Zealand are listed in table 1.

Of the Argids, the Ustollic Haplargids are predominant. Ustalfic Haplargids occur where there has been erosion. Aquic Haplargids occur on the lower slopes of fans. Irrigation can induce the low chroma colours required to meet the morphological criteria for Aquic Haplargids. The Natragids are common in the Tertiary sediment and schist saprolite land but occur only in small patches.

Of the Orthids, the Ustollic Camborthids are predominant. Lithic Camborthids occur on ridge and crest sites in basement rock land. Aquic Camborthids occur on the lower slopes of young fans. The extent of the Gypsiorthids is unknown. The Salorthids occur in minor areas in toes of young fans.

Taxonomic and Land Use Issues

Soil Moisture Regime at Subgroup Level

All NZ Haplargids and Camborthids fail the requirement for typic subgroups to be "dry in all

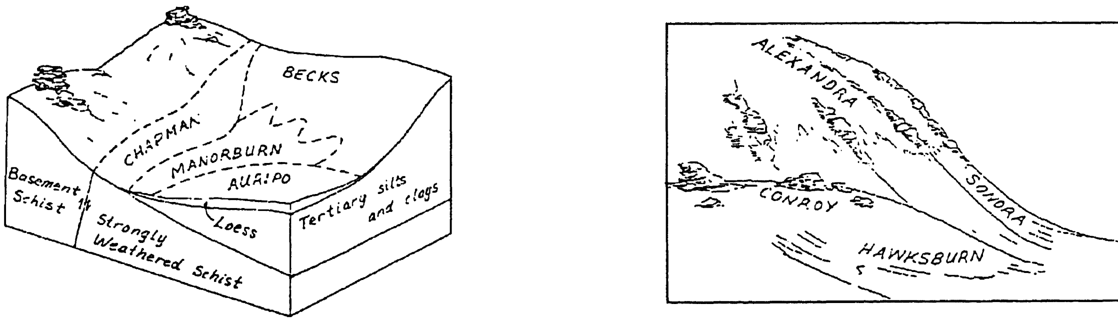


Figure 3. (a) Tertiary sediments and schist regolith land and (b) Basement rock land, showing soil series, their landforms, and classification. * m.m. = mixed mesic and i.m. = illititic, mesic

(a) **Chapman series, Ustollic Camborthid, Fine-loamy, m.m.**
Manorburn series, Typic Natragid, Fine-loamy, m.m.
Becks series, Ustollic Haplargid, Fine, i.m.
Auripo series, Ustollic Haplargid, Fine, i.m.

(b) **Alexandra series, Ustic Camborthid, Coarse-loamy, m.m.**
Sonora series, Ustollic Camborthid, Coarse-silty, m.m.
Conroy series, Lithic Camborthid, Coarse-loamy, m.m.
Hawksburn series, Ustollic Haplargid, Coarse-loamy, m.m.

TABLE 1. Aridisol subgroups occurring in New Zealand.

Suborder	Great Group	Subgroup	Example
Argids	Natrargids	Typic Natrargids	Manorburn
	Haplargids	Ustollic Haplargids	Waenga
		Ustalfic Haplargids	Hawksburn
		Aquic Haplargids	Ranfurly
Orthids	Gypsiorthids	Typic Gypsiorthids	Becks
	Camborthids	Ustollic Camborthids	Molyneux
		Lithic Camborthids	Conroy
		Aquic Camborthids	Linnburn
	Salorthids	Aquollic Salorthids	Linnburn salty

parts of the moisture control section for more than three-fourths of the time (cumulative) that the soil temperature is 5°C or more at a depth of 50 cm unless the soil is irrigated." The Newhall model indicates only 30% probability that soils will meet the typic requirement (Hewitt & Watt, in prep.).

Most Haplargids and Camborthids that are not severely eroded also fail the organic carbon requirement for typic subgroups. This leaves a choice between Ustollic and Xerollic subgroups which "... have an aridic moisture regime that borders on an ustic (or xeric) regime." The most appropriate assignment is unclear and hinges on the meaning of the words "borders on." If a geographic sense is intended, then an Ustollic subgroup is appropriate, because the Aridisols occur adjacent to soils with ustic moisture regimes. If a taxonomic sense is intended, however, then an objective decision cannot be made (a problem also noted by Thomas et al., 1981, page F12-9).

The soil moisture regime calculated for the Alexandra climate station is ustic-like to the extent that it meets (with 100% probability) the requirement for dryness 90 or more days cumulative each year. It is xeric-like to the extent that it meets (with 90% probability) the requirement for dryness in the 4 months following the winter solstice (with only 20% probability that the MCS will be moist). Although winter rainfall is particularly low, evapotranspiration is also very low. Moisture is stored, particularly in frozen topsoils, and becomes available for spring growth.

Until the most appropriate subgroup assignment can be clarified, the soils provisionally have been assigned to ustollic subgroups, because of their geographic association and their original grassland vegetation (a characteristic of ustollic subgroups noted by Guy Smith [Forbes, 1986]).

Argillic Horizons

The argillic horizon has proved to be a significant horizon in NZ Aridisols. Soils with argillic horizons are slowly permeable, have low air capacity (and therefore potential aeration problems under irrigation), high bulk density, and low macro-porosity. Clay contents vary from 12 to 37%. They are a significant root barrier with high penetration resistance, and, although available water capacities (10-1500 kPa) are moderate, only a small amount is readily available (10-100 kPa).

Exchangeable Na percent (ESP) values vary from 1 to 29%. Horizons with ESP of 15% or more fail the structural requirements of Natric Horizons.

Argillic horizons are easily recognized in the field in soils on Pleistocene land surfaces, because of their texture and clay coatings. A micromorphological study by Barrett (1971) confirmed the presence of argillans in two such profiles.

Field identification of argillic horizons in soils on Holocene land surfaces is less certain. Ped or pore surfaces have moist colour value of 4 or less with olive brown (2.5Y 4/4) being a common colour. The cutans, however, are too thin to observe with a 10x hand lens and often show a dull lustre rather than the waxy lustre expected for clay coatings. The micromorphology of these horizons has not been described. Particle-size measurements usually meet argillic horizon requirements, although parent material stratification is indicated in many profiles.

The soils have been provisionally assigned to the Haplargids because they have subsurface horizons with the accessory properties of argillic horizons. The identification problem has been handled in the NZ Soil Classification (Hewitt, 1989) by grouping together all soils with slowly permeable horizons that have cutans (whether they be argillans or suspected organic-iron-clay complexes).

Cambic Horizon

The recognition of cambic horizons is subjective in many NZ Aridisols. The soil parent materials are non-calcareous, and calcareous dustfall does not occur. Subsurface horizons containing calcium carbonate do occur as a product of weathering and leaching (Leamy and Rafter, 1972), and, in such soils, a calcium carbonate accumulation horizon establishes the lower boundary of the cambic horizon.

TABLE 2. Selected data for three NZ Aridisols

SOIL (LAB NO.)	HOR.	DEPTH TO BASE (cm)	CLAY %	CEC me%	ORG. C %
Lowburn (SB7584)	A	11	8	6.1	1.6
	Bw	41	4	4.0	0.3
	Bt	64	10	9.3	0.2
Waenga (SB9895)	A	28	14	8.5	1.9
	Bw	45	13	4.7	0.4
	Bt	69	21	8.5	0.3
Drybread (SB9892)	A	13	18	6.6	1.4
	AB	28	6	5.3	0.9
	Bt	67	12	9.5	0.3

In many soils on Holocene land surfaces, however, calcium carbonate horizons are absent, and chroma and hue do not change with depth. Cambic horizon recognition then must rely on "the presence of soil structure or absence of rock structure." Loess in Central Otago is non-calcareous, and even very young deposits do not show "rock structure" in the form of sedimentary layering. Furthermore, the soils without argillic horizons are commonly structureless-massive.

Permeability

Permeability, or field-assessed saturated hydraulic conductivity, is a particularly important quality for the interpretation of irrigable Aridisols. Under orchards using overhead sprinklers for frost fighting, the occurrence of slowly permeable layers requires either deep ripping or surface recontouring, to off-farm drainage outlets, and ridging of tree rows. The occurrence of rapidly permeable layers requires particular attention to water scheduling and lateral distribution of water.

Instead of reliance upon surrogate properties, "slowly permeable" and "rapidly permeable" diagnostic layers have been specified in the NZ Soil Classification (Hewitt, 1989). These layers are defined either by measured saturated hydraulic conductivity or by morphology, using a combination of pedality, particle size, and a semi-confined single vane-shear test (Griffiths, 1985).

In NZ Aridisols, slowly permeable layers include most argillic and natric horizons, as well as some cambic horizons. Rapidly permeable layers include most cambic horizons in soils with sandy-skeletal family particle size class.

Low Buffering Capacity

The clay fractions of NZ Aridisols are dominated by mica/illite (approx. 50-70%) and accompanied by kaolinite (approx. 10-20%). Minor amounts of smectite and interstratified chlorite-vermiculite occur in argillic horizons. Organic-carbon levels are low, and consequently the CEC is low, particularly in the most intensively used soils of the fans and terraces.

Mean oxalate-extractable iron and aluminium values are very low (0.2 and 0.06% respectively for all horizons). Consequently, phosphate retention (mean of 8% for all horizons) and absorbed sulfate (mean of 4 ug/g for A horizons) are also very low.

The soils therefore are poorly buffered and many chemical and physical properties are sensitive to management. Particular care needs to be taken in management of fertility and soil structure. The following land-use problems have resulted.

1. Inappropriate management advice can lead to the application of high rates of ammonium sulfate fertilizers as ground dressings under irrigated orchards. pH(H_20 soil: water = 1:4) levels have been depressed from around 7.0 to as low as 4.2, with subsequent toxicities, yield reduction, and increased disease susceptibility. This can be avoided by small and frequent applications of fertilizer, taking care to balance and monitor pH levels.
2. Fumigants also may depress pH. For example, one combined application of a fumigant and a nematicide caused a drop of 1 pH(H_20 soils water = 1:4) unit in A and B horizons.
3. Overhead sprinklers are used to counter spring frosts in stone fruit orchards. Topsoils become saturated and, with the impact of machinery, soil structure deteriorates, infiltration becomes slow, and aeration is greatly restricted. This also increases susceptibility to diseases such as bacterial blast.
4. Herbicide strips commonly are used to control grass and weed growth under orchards. Root studies indicate that herbicides penetrate topsoils and discourage root development in the upper 20 cm. This has particularly serious implications in sandy-skeletal soils where nutrients and available water are concentrated largely in topsoils.
5. Serious erosion has occurred when soils are cultivated in windy conditions. Dispersion and slaking measurements (McQueen, 1981) show very low aggregate stabilities in most soils.

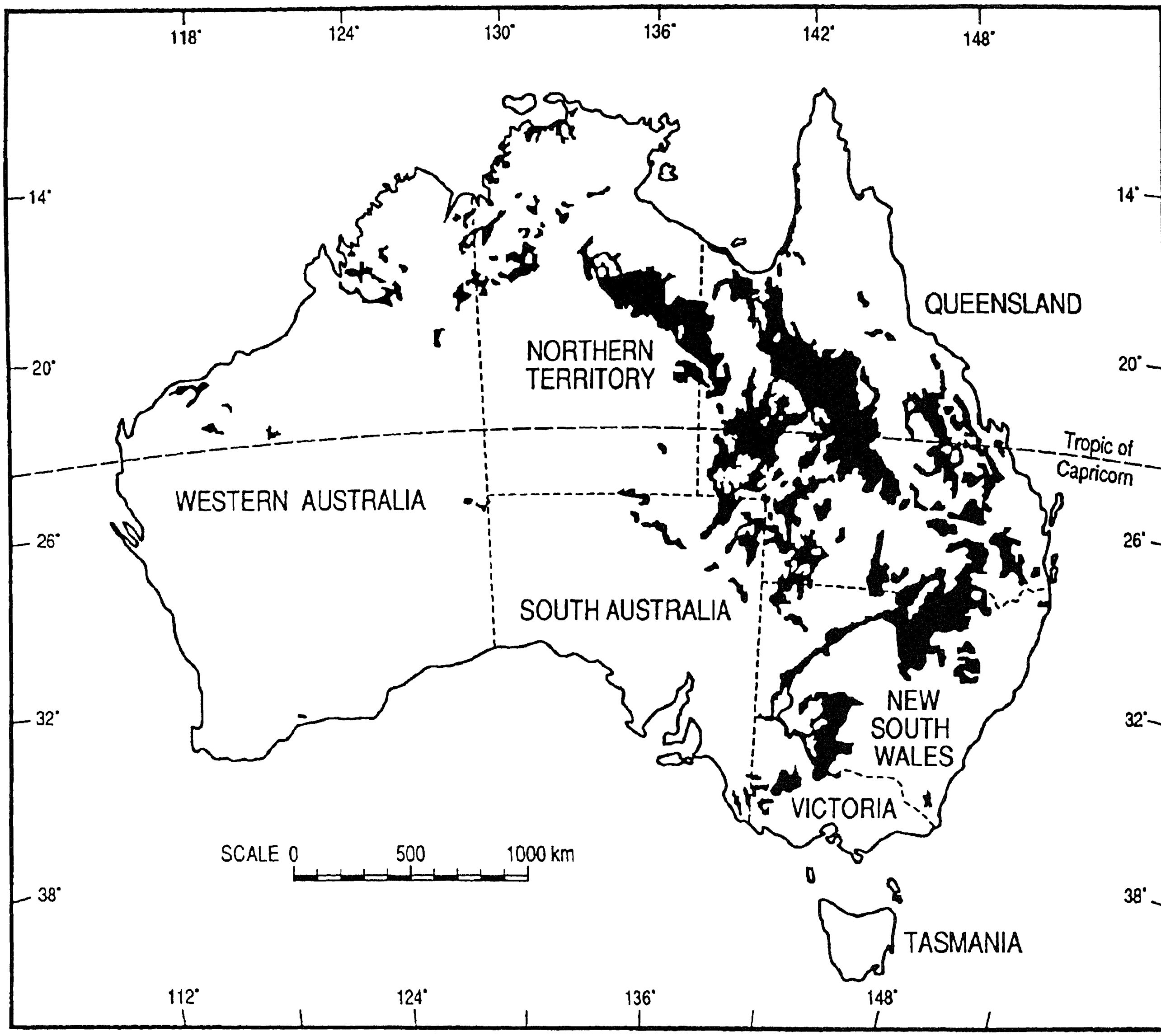


Fig. 1.Distribution of Vertisols in Australia (from Hubble et al., 1983).

sols occurs in the aridic soil moisture zone, with lesser ustic and xeric regimes and only very small areas of udic. In terms of median annual rainfall, the aridic areas range from about 100 to 650 mm, and ustic from as low as about 300 mm in the temperate south (winter dominant) to as high as 1800 mm in the tropics. Xeric range from about 300 to occasionally as high as 900 mm, while the few occurrences of Vertisols in udic regimes may be as low as about 1200 mm and as high as about 2000 mm. Much of the aridic and ustic regions are characterised by high rainfall variability.

Soil temperature regimes range from thermic in the south to hyperthermic over the greater part of the tropics. In spite of the relatively low latitudes in northern Australia, the isohyperthermic area is very small (Murtha, 1986).

Topography, Vegetation, and Soil Parent Materials

Australia is well known as a land of vast plains, and this is exemplified by most occurrences of Vertisols. In general, they occupy gently undulating plains with altitudes less than 500 m. Many occurrences are on wide flood plains of inland streams which receive their flood waters from higher rainfall regions in their headwaters. Flooding frequency varies greatly, with periods of up to a decade between major floods being not uncommon.

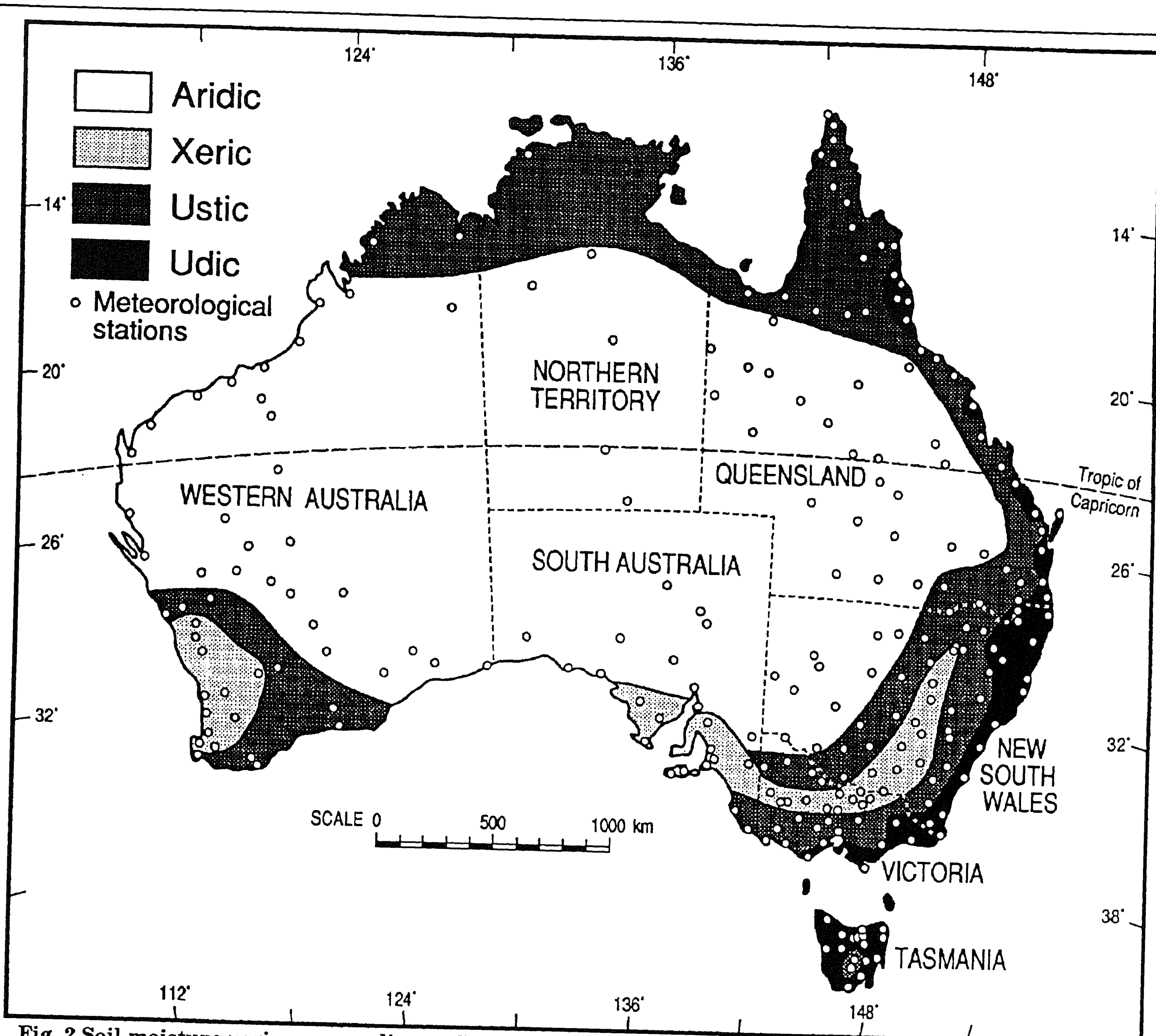


Fig. 2. Soil moisture regimes according to Soil Taxonomy, calculated from climatic data using a FORTRAN version of the USDA program MOREG 3.

The large semi-arid Vertisol regions are mainly grasslands, with the truly arid areas supporting only sparse low shrublands. In contrast, in eastern Queensland extensive areas of Vertisols originally supported Acacia forests.

As would be expected, the most common parent materials are alluvial and aeolian clayey sediments, sedimentary rocks such as shales, mudstones, and impure limestones, and basic igneous rocks — particularly, basalt.

Soil Morphology

Clear color and textural differences in the upper metre or so of the profile are not usual in most deep Vertisols, apart from the fairly common occurrence of a thin (up to 3 cm) surface crusty horizon of usually lower clay content. In Australia, soils with a thicker (> 3 cm) A horizon of lighter texture are not thought of as Vertisols but as texture-contrast soils, even though the top 18 cm may contain more than 30% clay after mixing.

There is a wide range in color, from red (most common in the arid and semi-arid regions) through shades of brown and grey to black. This color range is not necessarily related to rainfall or organic matter content, e.g., black soils derived from basalt may occur in semi-arid regions. Some grey or brown forms may become strongly red-mottled in their deep subsoils.

The soils have the characteristic Vertisol structure profile. Beneath various surface conditions (see below) is usually a subsurface hori-

zon of moderate, medium to coarse blocky or polyhedral peds, grading below 30-50 cm into a horizon of moderate, medium to coarse lenticular peds with prominent diagonal shear planes or inclined slickensides, features which extend to the full depth of the solum.

The natural condition of the surface soil varies widely in Australian Vertisols. These surface conditions do not necessarily relate to the climatic environment or the soil color, but cation status is involved and also probably clay content and mineralogy. The following classes can be distinguished in the virgin state, but insufficient evidence is yet available on whether these conditions always reform following disturbance such as cultivation.

(i)Massive or weakly structured surface crusty horizon _ 3 cm thick, often of lighter texture (lower clay content). This overlies pedal clay (blocky or polyhedral) which is not self-mulching.

(ii)After wetting and drying, a thin 5-10 mm surface flake forms which cracks into irregular polygons (plates) 3-10 cm diameter. These may be readily separated and removed from underlying pedal (blocky or polyhedral) clay which is usually not self-mulching.

(iii)Massive or weak, very coarse blocky A horizon, no surface flake forms on drying, and no surface crusty horizon.

(iv)Pedal A horizon (blocky or polyhedral) which is not self-mulching, no surface flake forms on drying, and no surface crusty horizon.

(v)Surface soil moderately to strongly self-mulching. Initial drying may form a very fragile thin (2-3 mm) surface seal which readily disintegrates to a self-mulch on further drying.

Carbonate in the form of discrete nodules or diffuse soft masses in the fine earth is a very common feature, while variable amounts of gypsum are almost ubiquitous in the more arid soils. Here also, a red-brown hardpan (similar to a duripan) may occur within or below the solum. Depth of solum varies enormously in the Australian Vertisols. Shallow forms (50 cm or less) are common where formed on underlying hard rocks, but also widespread are very deep forms with sola ranging up to 6 m or more. In between these extremes, depths of 1-2 m are usual.

Australia is unique in the variety and extent of gilgai associated with its soils. A summary has been given by Hubble et al. (1983). In brief, forms range from linear (long narrow parallel mounds and broader depressions on slopes, vertical interval < 30 cm and horizontal interval 5-8 m), through normal (small irregular mounds and subcircular depressions varying in size and spacing, vertical interval usually < 30 cm and horizontal usually 3-10 m), to melonhole gilgai (irregularly distributed large subcircular or irregular depressions, usually > 3 m greatest dimension, vertical interval 30 cm to as much as 2 m, horizontal 6-50 m). In most cases differences exist between profiles on mounds and depressions.

Physical, Chemical, and Mineralogical Properties

Physical Properties

Clay contents range from about 40 to 80%, the higher values usually in black soils derived from basic igneous rocks. Contents tend to be relatively uniform throughout the profile.

The other two major physical characteristics of Vertisols involve surface soil aggregate behaviour, and water infiltration, transmission, and storage properties. These are all interrelated and are dependent on clay content, mineralogy, and physico-chemical interactions. Recent Australian work in this very broad area has been reported by Williams (1983), Smith et al. (1984), and Coughlan et al. (1987).

The conceptual and practical difficulties of determining soil water storage in Vertisols both under rain-fed and irrigated conditions are well known. In Australia several approaches have been used. The concept of Plant Available Water Capacity (PAWC) has been summarised by Gardner (1985). PAWC is defined as the difference between the wet profile water content following irrigation and the dry profile water content under a stressed mature crop (sorghum) summed over the measured rooting depth. As such, PAWC reflects any limitation in the depth and degree of subsoil wetting and any physical and chemical constraints, including aeration and salinity, to subsoil root growth and water extraction. The data for a range of Vertisols listed by Gardner (1985) show PAWC values ranging from 70 to 140 mm over measured rooting depths of 40 to 120 mm (note that these rooting depths were not constrained by solum depth).

Other approaches to determining plant available water capacity of Vertisols have been the use of surrogate properties, such as regressions with cation exchange capacity and -1500 k Pa moisture, and the use of the soil chloride profile to estimate rooting depth, e.g., Shaw and Yule (1978), Gardner and Coughlan (1982), Ahern (1988), and Baker and Ahern (1989).

Williams (1983) points out that the recharge of a swelling clay under rainfall is a very different process from that under ponding, which is used in obtaining PAWC. 'Water entry capacity' therefore becomes all important, with surface soil conditions (including cracking and cultivation) and rainfall intensity dictating the amount of water that enters the soil. In short, under most dryland conditions the actual capacity of a Vertisol to store water is less important than the likelihood of a given amount of water actually to enter the profile and subsequently be available to the plant as stored water. Williams (1983) quotes values ranging from 60 to 260 mm of water stored under rainfall in Australian Vertisols.

An alternative, therefore, to the PAWC approach in determining soil water storage is to make use of the soil water retention properties in appropriate soil water balance models. While the shrink/swell properties of Vertisols can make for complex behaviour, Williams et al. (1983) have shown that the texture, structure, and mineralogy can be used to estimate the primary water retention functions. By using this approach, the soil properties are treated independently of the plant. This avoids some of the difficulties associated with the PAWC approach, which is determined by many interacting factors, including the water retention properties, the rooting depth and density, and the crop water use.

Chemical Properties

No comprehensive review is possible in this brief paper, and hence only some aspects will be considered.

As would be expected from the wide climatic range, organic carbon (and nitrogen) contents vary widely. Data in Spain et al. (1983) show for A horizons (0-5/15 cm depth) median organic carbon values of 2.2 and 0.8% for the black earths and grey, brown, and red clays respectively. Where the latter group of clays were under Acacia forest, the median value was 1.6%. Because the samples of the first two groups are from both virgin and cultivated sites, it is difficult to make meaningful comparisons with Vertisols elsewhere. For example, the dark Vertisols from India have mean organic carbon contents of about 0.6% (Sehgal and Sohan Lal, 1988), but these undoubtedly have been cultivated for much longer periods than soils in Australia.

The pH profiles of Australian Vertisols are of considerable interest and may well be unique. Three general classes occur: (i) alkaline - pH about 6.5 or more in the surface and increasingly alkaline with depth, (ii) alkaline/acid - pH about 6.5 or more in the surface but becoming strongly acid (pH 4.0 - 5.0) below depths of about a meter, (iii) acid - pH less than about 6.5 throughout, and usually strongly acid at depth. The alkaline/acid and most occurrences of the acid classes are widespread in deep and very deep grey and brown clays, often strongly gilgaied, in eastern Queensland under rainfalls ranging from about 500 to 700 mm.

Most soils in the arid and ustic environments are saline at depth, with some soils having high levels near the surface. Sodium chloride usually dominates the total salts (40 - 80%) except in soils containing large amounts of gypsum.

Cation exchange capacity (CEC) and exchangeable cations vary greatly — the former ranging from 20 to 80 c mol kg^{-1} — and obviously vary with clay content and clay mineral type. Calcium is usually the dominant exchangeable cation in the upper horizons and magnesium in the deeper subsoils. Many soils have appreciable sodium (ESP > 6) in the surface, rising to high values (30 or more) at depth. As a rule, the black earths are less saline and sodic than the others. The strongly acid subsoils are of interest, with base saturation values ranging from 60 to over 90% in spite of pH values of 4 to 5. Further study is required on the charge characteristics of these acid clays.

Clay Mineralogy

While the darker Australian Vertisols — particularly those derived from basalt — are usually smectite dominant, data in Norrish and Pickering (1983) and other unpublished data show that, of about 100 analyses of grey, brown, and red clays, over half are dominated by illite and kaolin, with some containing little or no smectite. In most cases soil parent material appears to play a predominant role in determining soil clay mineralogy. In the case of gilgai microrelief, there seems to be an inverse relationship between smectite content and magnitude of gilgai development, particularly exemplified in the case of melonhole gilgai.

Land Use Aspects

In much of the arid and semi-arid regions, Vertisols are used for grazing of native pastures by sheep and cattle. Stocking rates are very low in the arid areas. In central eastern Queensland most of the original Acacia forests have been cleared and sown to improved grass pastures. Vertisols are extensively used for dry land agriculture in the east and south, usually between rainfall limits of about 300 and 800 mm. A wide range of winter crops (mainly wheat, safflower, barley, and oats) and summer crops (chiefly sorghum, maize, cotton, soybeans, sunflower, and millets) are grown. Wheat is grown as far north as 22°S latitude. Vertisols also are extensively irrigated, the chief areas being central and south Queensland and northern New South Wales (cotton with some grain and fodder crops and small areas of rice in the north) and large areas on the Riverine Plain in south-eastern Australia, where the main crop is rice.

In the case of irrigation, most Vertisol problems involve restricted water entry and transmission caused by adverse soil physical conditions, which in turn are at least partly induced by high sodium in the upper part of the profile. These conditions are very common in the irrigation areas on the Riverine Plain. In particular, gypsum has been used extensively as a soil ameliorant, and the recognition of gypsum-responsive soils has been widely studied (Loveday 1974).

In northern New South Wales and Queensland, rain-fed cropping on Vertisols is largely constrained by soil water, and farmers have long used the practice of growing one crop per year and fallowing for water storage during the off season. Gardner et al. (1988) have pointed out that fallowing is but one of three strategies to optimize production on Vertisols in a semi-arid environment of high rainfall variability. The other two methods are (i) using rain when it falls by opportunity cropping, provided the soil profile water stored equals or exceeds some specified amount. (The rationale behind this is to reduce evaporation losses from fallow land. This strategy involves either winter or summer or double cropping each year); (ii) matching cropping to plant available water capacity (PAWC) and climate by means of cropping models. The presence of a potentially large PAWC is one thing, but other important questions are how often matching can be achieved and what are the probabilities of adequate planting and growing season rains. Cropping models thus can indicate the probability of achieving high yields and the chances of avoiding low yields resulting in monetary loss.

Other soil-related land use problems include various degradation problems such as erosion, structural decline, plough pan formation, and organic matter decline, all of which are serious to varying degrees.

Classification

Soil Taxonomy

Since the advent of *Soil Taxonomy*, the term Vertisol has wide currency in Australia, although subdivisions below the order level are seldom used. This is largely because the soil moisture-based suborders are not realistic in a land use sense, nor does the present 'pell' and 'chrom' separation reflect any meaningful or consistent distinction.

Recent proposed changes to the classification of Vertisols in *Soil Taxonomy* (Comerma et al., 1988) are mainly at the great group and subgroup level, so that apart from the suggested addition of Aquerts and Borerts, the present suborders remain unchanged. In the Australian context this is unsatisfactory for several reasons. From a land use point of view there are no differences between 'wetter' Torrerts and 'drier' Usterts and Xererts, and certainly in eastern Australia crops are consistently grown in areas thought to be aridic.

A major weakness at the suborder level is the present use of cracking patterns as a surrogate for moisture regimes. No actual measurements of this have ever been undertaken in Australia, and it seems unlikely that definitive studies have been undertaken elsewhere in the world. Further, as Dudal and Eswaran (1988) have pointed out, cracking is influenced by a number of soil properties other than climatic conditions; this is very evident in the wide range of Vertisols occurring in the Australian arid zone.

Cracking patterns aside, if the concept of soil moisture regimes is to be used at the suborder level in Vertisols (or any other order for that matter), better methods of defining and estimating them must be devised. Some of the major problems of the present MOREG 3 method have been discussed by Isbell and Williams (1981). Given the arguments in that paper about the shortcomings of the soil moisture control section concept, and the somewhat unique hydrological

properties of Vertisols, it seems that a new approach is required. The most obvious is the use of improved water balance models which are currently under development, a suggestion also made recently by Bouma and Loveday (1988).

Proposed changes at the great group level (Comerma et al., 1988, and unpublished ICOMERT Circular Letter No. 5) include great groups based on the presence of salic, calcic, and gypsic horizons in the 'torric' and 'ustic' suborders. These would probably represent relatively minor soils in Australia in 'ustic' regions, but would be more common in the more arid regions, particularly gypsic soils. The proposal in Circular Letter No. 5 for an acid great group of Usterts and Uderts would cater to some but not the majority of Australian Usterts which have acid subsoils. In summary, the great majority of Australian Vertisols would fall into the proposed 'Hapl' great groups.

In an Australian context, the proposals of interest at the subgroup level include the use of ESP and the suggestion in Circular Letter No. 5 of a separation of dark forms based on value 3 or less and chroma 2 or less. However, the proposals for further subdivision of implied soil moisture regimes based on cracking frequency is of course subject to the same deficiencies as discussed earlier in this paper.

Australian Classification

The traditional usage in Australia of the terms black earths and grey, brown, and red clays has some utility in that it tends to separate the generally more fertile and less variable dark forms from all the rest. In the *Factual Key* (Northcote, 1979), a similar high level separation based on color was used, with some attention paid to surface soil condition and solum depth at lower hierarchical levels.

In a proposed new classification currently under development, the Vertisol equivalents are defined in a similar fashion (but with a lower clay limit of 35%, no mixing criterion, and no depth restriction). The first suborder to key out contains the 'aquic' soils — those that are saturated in at least some part of the upper 0.5 m continuously for prolonged periods in most years. The remaining suborders are based on color, using generalized groupings based on the Munsell chart giving classes termed red, brown, grey, and black. The rationale for this is the usual higher clay and smectite content and fertility levels of the dark forms (mainly due to their basic parent materials). In the case of the other colors, a tentative separation has been made on the basis of their distinctive appearance. Justification of this subdivision will depend on the result of searching the extensive Australian soil databases.

Great groups are in general distinguished on the nature of the soil surface (some or all of the classes listed earlier). Subgroup criteria include presence of a red-brown hardpan (duripan), a bleached A2 horizon, carbonate and/or gypsum, exchangeable sodium percentage, and presence of acid horizons. The usual difficulties arise in defining subgroups, because few criteria are mutually exclusive. At the family level, differentiae presently used include depth of solum and clay content. The use of a CEC/clay ratio as a partial surrogate for clay mineralogy is being investigated.

Literature Cited

Ahern, C.R. 1988. Comparison of models for predicting available water capacity of Burdekin soils, Queensland. Aust. J. Soil Res. 26:409-423.

Baker, D.E., and C.R. Ahern. 1989. Estimates of effective rooting depth for predicting available water capacity of Burdekin soils, Queensland. Aust. J. Soil Res. 27:439-454.

Bouma, J., and J. Loveday. (1988). Characterizing soil water regimes in swelling clay soils. p. 83-96. *In* L.P. Wilding and R. Puentes (ed.) Vertisols: their distribution, properties, classification and management. Technical Monograph No. 18, Soil Management Support Services.

Comerma, J.A., D. Williams, and A. Newman. 1988. Conceptual changes in the classification of Vertisols. p. 41-54. *In* L.P. Wilding and R. Puentes (ed.) Vertisols: their distribution, properties, classification and management. Technical Monograph No. 18, Soil Management Support Services.

Coughlan, K.J., D. McGarry, and G.D. Smith. 1987. The physical and mechanical characterization of Vertisols. p. 89-105. *In* M. Latham and P. Ahn (ed.) Management of Vertisols under semi-arid conditions. IBSRAM Proceedings No. 6.

Dudal, R., and H. Eswaran. 1988. Distribution, properties and classification of Vertisols. p. 1-22. *In* L.P. Wilding and R. Puentes (ed.) Vertisols: their distribution, properties, classification and management. Technical Monograph No. 18, Soil Management Support Services.

Gardner, E.A. 1985. Assessment of the plant available water capacity of swelling soils proposed for irrigation development. p. 67-79. *In* Planning and management of water for agriculture in the tropics. Proc. Fifth Afro-Asian Regional Conference, Townsville. Int. Comm. on Irrigation and Drainage.

Gardner, E.A., and K.J. Coughlan. 1982. Physical factors determining soil suitability for irrigation crop production in the Burdekin-Elliot River area. Technical Report No. 20, Agricultural Chemistry Branch, Dept. Primary Industries, Brisbane.

Gardner, E.A., K.J. Coughlan, and D.M. Silburn. 1988. Soil water measurement and management on Vertisols in Queensland, Australia. p. 131-165. *In* S.C. Jutzi, I. Haque, J. McIntire and J.E.S. Stares (ed.) Management of Vertisols in sub-Saharan Africa. Proc. Conference held at ILCA, Addis Ababa, Ethiopia 1987.

Hubble, G.D. 1984. The cracking clay soils: definition, distribution, nature, genesis and use. p. 3-13. *In* J.W. McGarity, E.H. Hoult and H.B. So (ed.) The properties and utilization of cracking clay soils. Reviews in Rural Science No. 5. Proc. Symp. Univ. New England, Armidale, 1981.

Hubble, G.D., R.F. Isbell, and K.H. Northcote. 1983. Features of Australian Soils. p. 17-47. *In* Soils: an Australian viewpoint. CSIRO, Melbourne/Academic Press, London.

Isbell, R.F., and J. Williams. 1981. Dry soils of Australia: characteristics and classification. p. 124-150. *In* F.H. Beinroth and A. Osman (ed.) Proc. Third Int. Soil Classification Workshop, ACSAD, Damascus.

Loveday, J.L. 1974. Recognition of gypsum-responsive soils. Aust. J. Soil Res. 25:87-96.

Murtha, G.G. 1986. Soil temperature regimes - a tropical experience. p. 23-27. *In* A.W. Moore (ed.) The USDA soil taxonomy in relation to some soils of eastern Queensland. CSIRO Aust. Div. Soils, Divl. Rep. No. 84.

Norrish, K., and J.G. Pickering. 1983. Clay minerals. p. 281-308. *In* Soils: an Australian viewpoint. CSIRO Melbourne/Academic Press, London.

Northcote, K.H. 1979. A Factual Key for the recognition of Australian soils. 4th edn. Rellim, Glenside, S.A.

Probert, M.E., I.F. Fergus, B.J. Bridge, D. McGarry, C.H. Thompson, and J.S. Russell. 1987. The properties and management of Vertisols. IBSRAM/CAB International, Wallingford, U.K.

Sehgal, J.H., and J.C. Bhattacharjee. 1988. Typic Vertisols of India and Iraq - their characterization and classification. Pedologie XXXVIII:67-95.

Sehgal, J.L., and Sohan Lal (ed.) 1988. Benchmark swell-shrink soils of India. National Bureau of Soil Survey and Land Use Planning, Nagpur, India.

Shaw, R.J., and D.F. Yule. 1978. The assessment of soils for irrigation, Emerald, Queensland. Technical Report No. 13, Agricultural Chemistry Branch, Dept. Primary Industries, Brisbane.

Smith, G.D., D.F. Yule, and K.J. Coughlan. 1984. Soil physical factors in crop production on Vertisols in Queensland, Australia. p. 87-104. *In* E.T. Craswell and R.F. Isbell (ed.) International Workshop on Soils, Townsville 1983. Australian Centre for International Agricultural Research.

Spain, A.V., R.F. Isbell, and M.E. Probert. 1983. Soil organic matter. p. 551-563. *In* Soils: an Australian viewpoint. CSIRO, Melbourne/Academic Press, London.

Stace, H.C.T., G.D. Hubble, R. Brewer, K.H. Northcote, J.R. Sleeman, M.J. Mulcahy and E.G. Hallsworth. 1968. A handbook of Australian soils. Rellim, Glenside, S.A.

Williams, J. 1983. Soil hydrology. p. 507-530. *In* Soils: an Australian viewpoint. CSIRO, Melbourne/Academic Press, London.

Williams, J., R.E. Prebble, W.T. Williams, and C.T. Hignett. (1983). The influence of texture, structure and clay mineralogy on the soil moisture characteristic. Aust. J. Soil Res. 21:15-32.

Properties and Classification of Cold Aridisols in Montana

C. Wang and Tom Keck, Jerry Nielsen, Robert Richardson, and Gordon Decker[1]

Abstract

Proposed changes in the Aridisol order would have a profound affect on the future classification of most Cold Aridisols in Montana. Aridisols are largely confined to semi-arid regions of the state which receive between 254 and 356 mm (10 to 14 inches) of mean annual precipitation. They are separated from Aridic subgroups of Mollisols in these areas because they fail to meet the color requirements for a mollic epipedon. Organic carbon, crop yield, and soil climate data were examined as they relate to the current and future classification of Aridisols in Montana. Surface horizons of both Aridisols and Mollisols, within the semi-arid region, had similar amounts of organic carbon. Crop yield data were limited for Montana soils. The data that were available did not support the hypothesis that the darker surface color of Mollisols represented a substantially higher yield potential. The Montana Agricultural Potentials System (MAPS) was used to evaluate the geographic distribution of soil climatic variables across the state.

Introduction

Aridisols have been mapped extensively throughout semi-arid regions of Montana. They account for a major portion of the taxa used in the eastern half of the state and in the intermountain valleys of southwestern Montana. Aridic soil moisture regimes are implied by the classification; however, the soil moisture regimes for most of these areas can be more accurately described as Aridic intergrades of the ustic moisture regime. Water is limiting to the growth of mesophytic plants but not to the extent typically associated with Aridisols. Sufficient spring moisture allows for successful crop-fallow operations in many of these areas.

This paper examines the available climatic, organic carbon, and crop yield data for cold Aridisols as they have been mapped in Montana. The underlying question is whether these soils are most accurately classified as Aridisols or would be more appropriately assigned to Inceptisol and Alfisol soil orders. Proposed changes in the Aridisol order, specifically the allowance for Aridisols with a mollic epipedon, have created a need to more accurately identify aridic, ustic, and xeric soil moisture regimes in Montana.

Soil Climate

The majority of Aridisols in Montana have been mapped in areas having a frigid soil temperature regime. Three of the four Montana Aridisol sites on the cold Aridisol tour have a frigid temperature regime. Mesic Aridisols have been mapped in several counties along the Yellowstone River in the south-central part of the state. The Stormitt series is the lone representative of a mesic Aridisol in the Montana portion of the tour. To date, no cryic Aridisols have been mapped. The potential exists for identifying such cryic Aridisols in the intermountain valleys of southwestern Montana, specifically in Beaverhead county (Nimlos and Tippy, 1981).

Soil series often are separated by precipitation zones. In Montana, most soils mapped as Aridisols occur within a 254 to 356 mm (10 to 14 inch) mean annual precipitation range (Fig. 1). Areas receiving less than 254 mm of mean annual precipitation are limited in the state to isolated pockets in the rainshadow of mountains. These occur south of Bridger along the Clark's Fork of the Yellowstone River and near Dillon along the Beaverhead and Jefferson Rivers (Fig. 1). Aridisols generally have not been mapped in any areas receiving more than 356 mm of mean annual precipitation.

Classification

The classification of Aridisols in Montana would undoubtedly change due to the proposed changes in the Aridisol order. As things stand today, all Aridisols in the state, having a frigid temperature regime and occurring within the 254 to 356 mm (10 to 14 inch) precipitation zone, are in Borollic subgroups. Typic subgroups are restricted to areas receiving less than 254 mm of mean annual precipitation.

[1]Graduate Research Assistant and Professor of Soil Science, Dep. of Plant and Soil Science, Montana State Univ., Bozeman, MT 59717; State Soil Correlator, and State Soil Scientist, Soil Conservation Service, USDA, Bozeman, Montana, 59715.

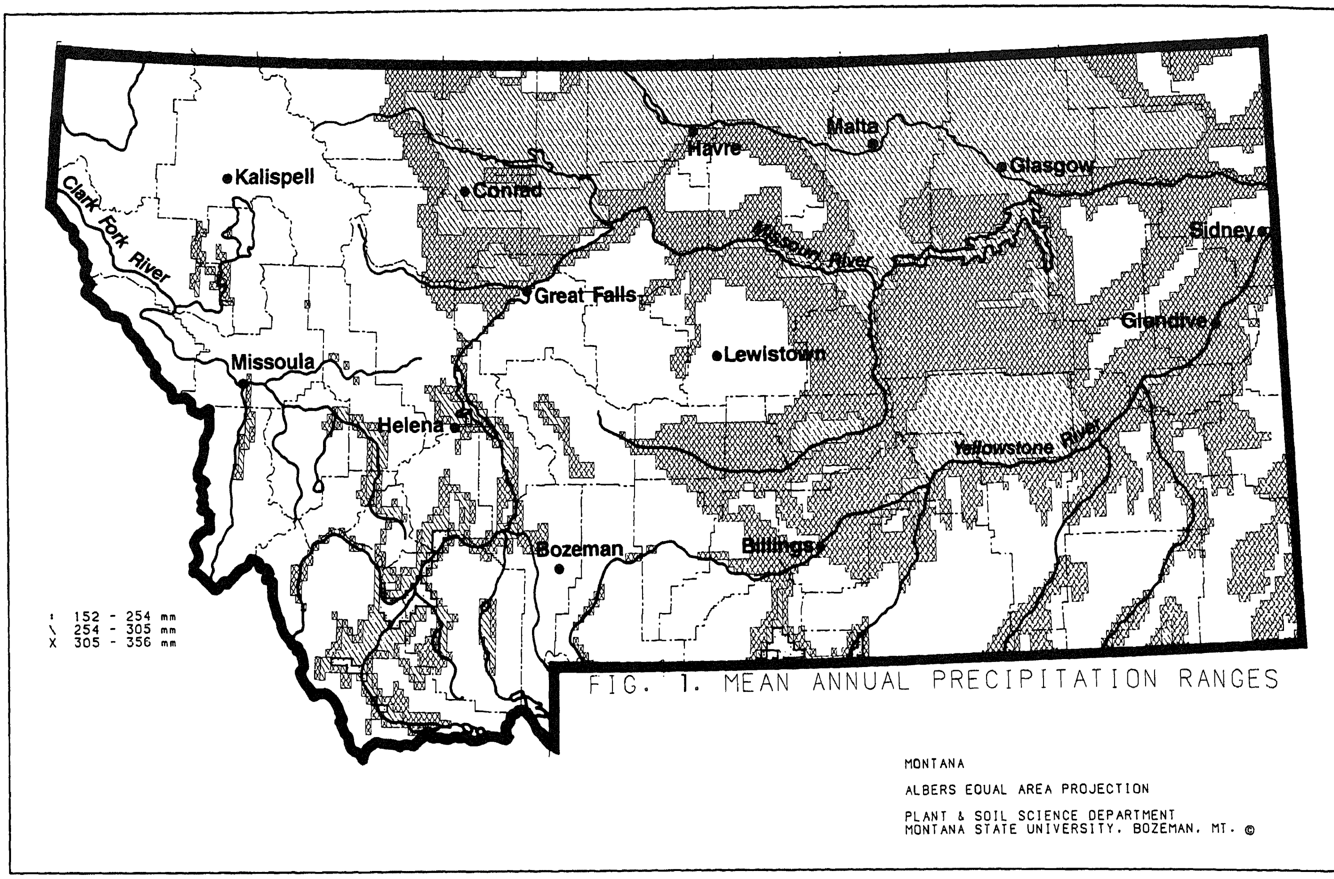


FIG. 1. MEAN ANNUAL PRECIPITATION RANGES

Most 'mesic' Aridisols are classified as Ustollic, Ustertic, Haplustollic, and Glossic Ustollic subgroups.

The Argids currently mapped in Montana are Natrargids and Haplargids with some Paleargids. Camborthids and Calciorthids comprise most of the Orthids. Two of the Aridisol tour sites in Montana have Camborthids and two have Calciorthids. To the best of our knowledge, only one Gypsiorthid series and two Salorthid series have been classified and mapped in Montana.

Changes in Classification

The proposed classification changes will have different effects on taxa as they are currently mapped in Montana. For example, most Borollic Haplargids will remain as Borollic Haplargids based on Version 6.0 of the revised Aridisol keys (International Committee on Aridisols, 1989). Most Borollic Natrargids and Paleargids would change to Haplic and Ustalfic subgroups, respectively. This is because in the revised criteria Borollic subgroups of these soils have a color value, when crushed, darker than 3.5 moist and 5.5 dry. Most Ustollic Haplargids would switch to Ustalfic Haplargids and most Ustollic Natrargids would probably fit within the Haplic Natrargid classification. Above the subgroup level, the classification of Paleargids and Natrargids would be unchanged. Some Haplargids would change to Calciargids.

The Camborthids would mostly become Haplocambids in the revised classification. The Calciorthids would primarily be Haplocalcids with a few Petrocalcids. The Borollic subgroup would remain with the Haplocalcids but would change to Ustocreptic for the Haplocambids. The current Ustollic subgroup of Camborthids would be classified as Ustochreptic Haplocambids.

These projected changes are based on the overall revisions in Version 6.0 of the International Committee report on Aridisols. They are not based on any detailed study of specific series. In some cases, the proposed revisions would inevitably create inconsistencies with the way a given series has been mapped in the past. Our assumption in the above projected changes is that these soils would remain within the Aridisol order.

Relation to Other Soil Orders

Aridisols have been mapped in close association with Vertisols, Entisols, and Mollisols

throughout semi-arid areas of Montana. Vertisols are separated from Aridisols by cracks 1 cm or wider to a depth of 50 cm, and by slickensides close enough to intersect, or by wedge-shaped natural structural aggregates that have their long axes tilted 10° to 60° from the horizontal. Entisols are separated from Aridisols by their lack of diagnostic horizons beyond an ochric epipedon. The Aridic subgroups of Mollisols are separated from Aridisols on the basis of soil color of the surface horizon or horizons. The Aridic subgroups of Mollisols would be included within the Aridisol order, based on the proposed revisions. This change is because the revisions allow for mollic surface colors (i.e., mollic epipedons) in Aridisols and because Aridisols key out before Mollisols in the key to soil orders. The Scobey series, included at one of the Aridisol tour sites, is an example of an Aridic Argiboroll that would need to be reclassified as an Aridisol if the area was in fact aridic.

Inceptisols and Alfisols have generally been restricted in Montana to areas receiving more than 356 mm of mean annual precipitation. Thus, these soil orders have not generally been mapped in close association with Aridisols. A decision was made, when *Soil Taxonomy* was implemented, that Alfisols in Montana could not occur under grassland vegetation. This decision meant that grassland soils, having an argillic horizon but not meeting the color requirements for a mollic epipedon, had to be classified as Aridisols. Also, there were no provisions in *Soil Taxonomy* for aridic intergrades of Inceptisols with a frigid temperature regime. For these reasons, the Aridisol order was used extensively in Montana for areas that have aridic intergrades of the ustic soil moisture regime.

As a result, aridic intergrades of Mollisols (i.e., Aridic Argiborolls) with an ustic soil moisture regime are mapped in the same precipitation zone with ustic intergrades of Aridisols (i.e., Borollic Camborthids). Mapping of Aridisols today in semi-arid areas of Montana is still influenced by earlier decisions.

The proposed changes in Aridisol taxonomy provide Montana with an incentive and an opportunity to re-evaluate the soil moisture regime for most Aridisols within the state which receive between 254 to 356 mm (10 to 14 inches) of mean annual precipitation. Potential use and management should be of primary concern in decisions about the future classification of Aridisols in Montana.

Soil Organic Matter

Hans Jenny, in 1930, described the basic relationship under grassland vegetation between soil organic matter and climate. Organic matter increases with decreasing temperature and with increasing moisture (Jenny, 1930). Mollisols occurring in moister and cooler environments contain considerably more organic carbon than do Aridisols in hotter and drier climates. However, this may not be the case when both orders occur within the same soil temperature and moisture regime. The assumption is that the dark surface color of Mollisols represents higher organic carbon levels in the soil. The mollic epipedon has been viewed as a "fossil record" of past root growth and a predictor of future production (Cannon and Nielsen, 1984). Yet if Jenny's conceptual model of soil organic matter in grasslands is correct, we might expect similar ranges in organic carbon for both Aridisols and Mollisols when comparing the two within the same temperature and moisture zones.

Individual pedons will vary in organic carbon content for a host of site specific reasons: topographic position, parent material, soil texture, and management history (Nichols, 1984; Franzmeier et al., 1985; Sims and Nielsen, 1986; Aguilar et al., 1988; Yonker et al., 1988). Acknowledging this variation, we question how well surface color of Mollisols, versus Aridisols, in semi-arid areas of Montana, represents differences in the amount of organic carbon in surface horizons.

Organic Carbon Data for Montana Soils

We used the data available in the Montana Pedon Database (Decker, 1972) to compare organic carbon levels between Mollisols and Aridisols mapped within the 254 to 356 mm (10 to 14 inch) precipitation zone in Montana. These data originated from two sources, SCS soil characterization data from the National Soils Lab in Lincoln and a statewide BLM study of range and soil characteristics in Montana (McDaniel et.al., 1982). The SCS data represents both cropland and rangelands sites, while the BLM data is for rangeland sites only. Each data set was handled separately to compare results from different sources. Average organic carbon levels were calculated for the top 0 to 18 cm (0 to 7 inches) and for 18 to 36 cm (7 to 14 inches) depths for all pedons, and comparisons were made based on taxonomic classifications (Fig. 2 and 3).

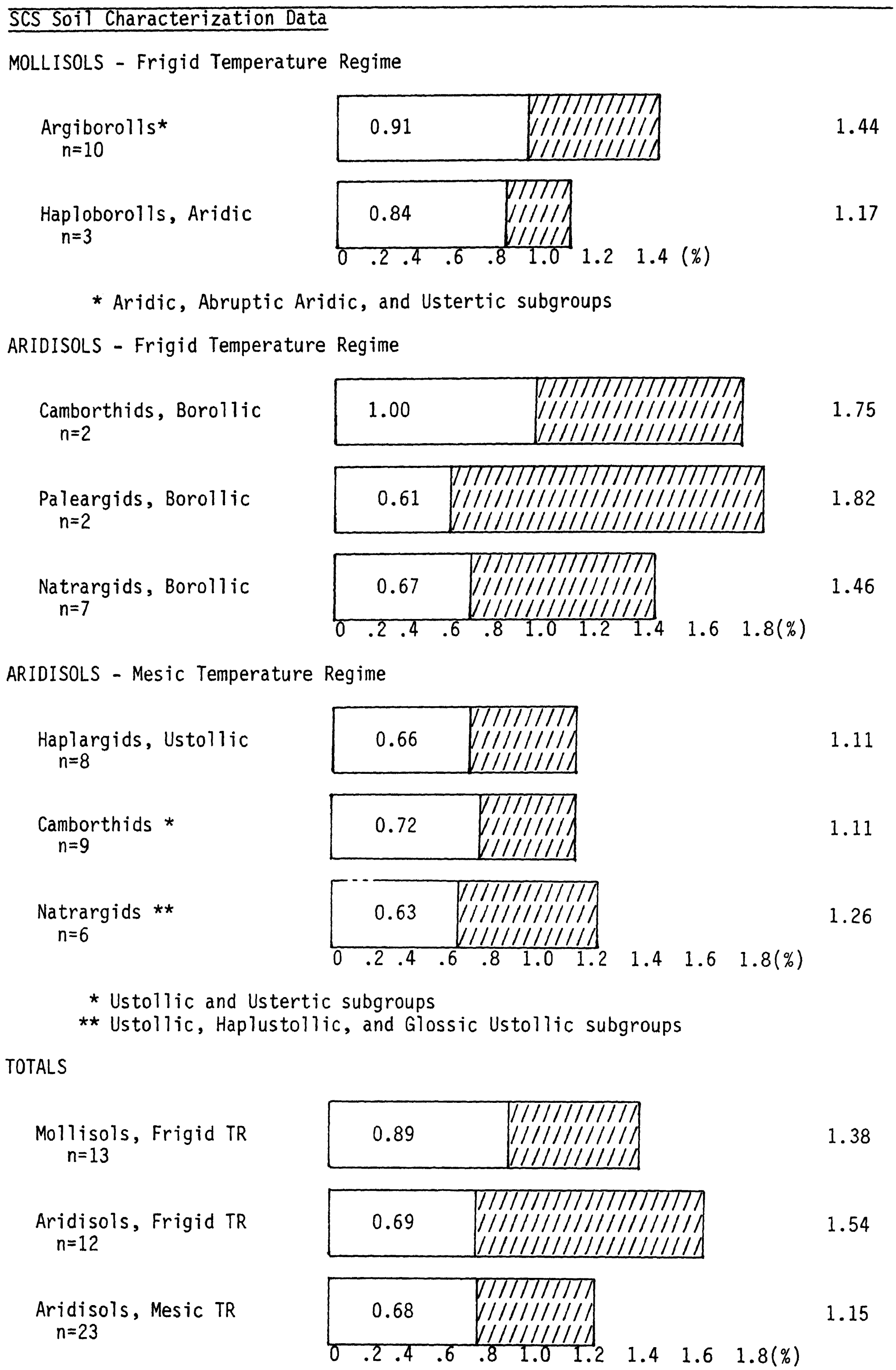


Figure 2. Average organic carbon content (%) in the surface 18 centimeters (numbers to right of bar graphs) and in the 18 to 36 cm depth (numbers within left side of bar graphs) of Mollisols and Aridisols mapped within the 254 to 356 mm precipitation zone in Montana. SCS soil characterization data (Decker, 1971).

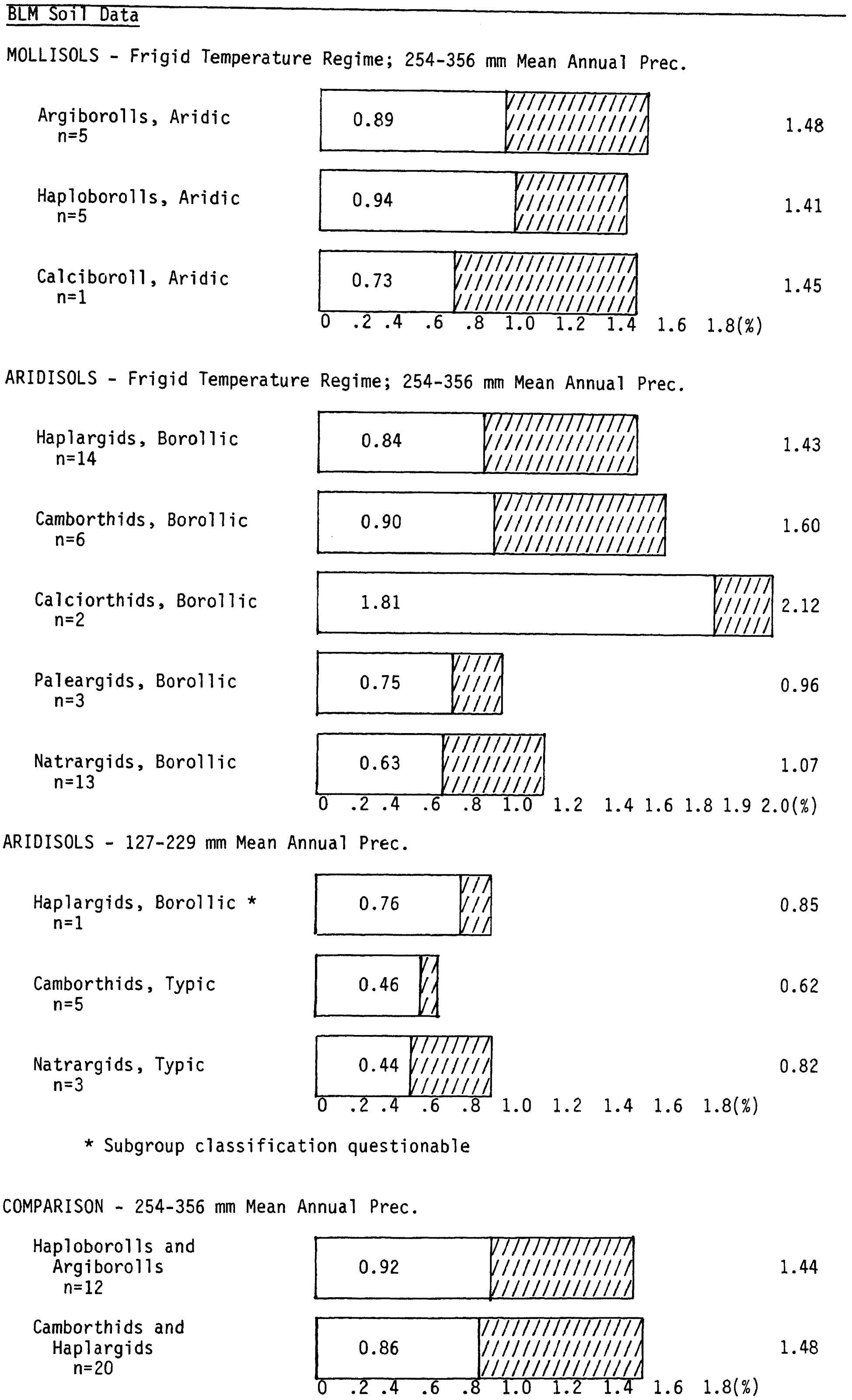


Figure 3. Average organic carbon content (%) in the surface 18 centimeters (numbers to right of bar graphs) and in the 18 to 36 cm depth (numbers within left side of bar graphs) of Mollisols and Aridisols mapped in arid and semi-arid regions of Montana. BLM soil characterization data (McDaniel, 1982).

SCS Data

Looking first at the SCS Data, there were only two Borollic Camborthid and two Borollic Paleargid pedons in the database with organic carbon data, but the average organic carbon levels in the top 18 cm for these groups were 1.75% for the Borollic Camborthids and 1.82% for the Borollic Paleargids. The sodium affected Borollic Natrargids, based on seven samples, had an average organic carbon content in the top 18 cm of 1.46 percent. Aridisols with a frigid temperature as a whole had an average organic carbon content of 1.54% in the top 18 cm and 0.67% in the 18 to 36 cm depth.

In contrast, Mollisols, within the same precipitation zone and with the same frigid soil temperature regime, had an average organic matter content of 1.38% in the top 18 cm and 0.89% in the 18 to 36 cm depth. Averages for the top 18 cm in different subgroups were 1.50% for Aridic Argiborolls, 1.40% for Abruptic Aridic Argiborolls, 1.17% for Aridic Haploborolls, and 1.39% for Ustertic Argiborolls. Admittedly, this initial data set is small, but at first glance, the data do not support the idea that Mollisols have higher levels of organic carbon than associated Aridisols in the same climatic zone.

There was an abundance of SCS Data for Aridisols with a mesic soil temperature regime. If Jenny's conceptual model is correct, then warmer temperatures should result in lower organic carbon levels. McDaniel and Munn (1985) found that organic matter did decrease in both Mollisols and Aridisols as temperature became warmer. The SCS data substantiate this observation. The average for 23 mesic Aridisol sites was 1.15 percent for the top 18 cm and 0.68% for the 18 to 36 cm depth (Fig. 2).

BLM Data or (Bureau of Land Management Data)

Care must be taken in drawing any conclusions from such a limited data set. The BLM data provided an excellent opportunity to check for similar trends in a completely separate data set. Again, comparisons were made based on subgroups and average organic carbon levels calculated for the top 18 cm and then for the 18 to 36 cm depth.

Borollic Camborthids in the BLM data had an average organic carbon content of 1.6% (n = 6) in the top 18 cm. The high for this group was 2.26%; the low was 1.06%. Borollic Haplargids had an average in the top 18 cm of 1.43% (n = 14) with a high of 2.05% and a low of 0.97%. Borollic Calciorthids, based on a sample of only two, had an average of 2.12% organic carbon in the top 18 cm; the high was 2.28% and the low was 1.95%. The corresponding organic carbon averages, in the top 18 cm, for the Aridic subgroups of Mollisols were: 1.41% (n = 7) for Aridic Haploborolls (high 2.11% and low .71%), 1.48% (n = 5) for Aridic Argiborolls (high 1.97% and low .99%), 1.45% for the only Aridic Calciboroll. Averages for the 18 to 36 cm depth were also similar between the above groups of Mollisols and Aridisols, except for the unusually high organic carbon levels in the two Borollic Calciorthids.

The BLM data, unlike the SCS characterization data, show consistently low levels of organic carbon in the top 18 cm of Borollic Natrargids (avg. = 1.07%, n = 13) and Borollic Paleargids (avg. = .96%, n = 3). The low average for Natrargids is expected due to reduced grass production on sodium and salt-affected soils. The low average for the Paleargids is surprising. It may be explained in part by the borderline classification, Borollic Paleargid/Borollic Natrargid, for two of the three samples. It seems reasonable to conclude that sodium and salt-affected Aridisols will have reduced organic carbon levels.

The BLM data set also included eight Aridisol sites from areas receiving 127 to 229 mm (5 to 9 inches) of mean annual precipitation. As expected, the drier Aridisols had consistently lower organic carbon levels than either Mollisols or Aridisols receiving 254 to 356 mm of mean annual precipitation.

Conclusions from Available Organic Carbon Data

The data available supports Jenny's initial idea that soil organic matter varies primarily in response to temperature and moisture. Borollic Aridisols and Aridic Mollisols have comparable levels of organic carbon in their surface horizons despite differences in surface colors. Warmer soils, as in mesic Aridisols, or drier soils, as in Typic Aridisols, have correspondingly lower levels of organic carbon. Soil factors, such as salinity or sodicity, which reduce plant production will also reduce organic carbon levels. Soil color is an attribute useful for separating soil series in the field. It does not necessarily, however, represent consistent differences in organic carbon levels between Mollisols and Aridisols in semiarid areas.

Humic - Fulvic Acids

Differences in soil color between Aridisols and Mollisols in the same precipitation zone could be related to differences in their ratios of humic and fulvic acids. Extracted humic acid is black or dark brown. Fulvic acid, on the other hand, is normally tan or lighter brown. The transition from wetter to drier grassland soils not only reduces the amount of grass and therefore, organic carbon production, but also reduces the intensity of the humification process (Anderson, 1979). Not only is there less organic carbon, but also a higher proportion of it is the lighter colored fulvic acid. Shields et al (1968) found higher than expected organic carbon contents in Gray Wooded Soils (Alfisols) because they contained a higher proportion of fulvic acids than the Chernozemic soils (Mollisols).

Crop Yield Data

Site specific crop yield data are extremely limited for most of Montana's agricultural soils. An on-going cooperative crop yield study by the Soil Conservation Service and Montana State University is intended to determine how dryland crops respond to different soil, climate, and management factors. Table 1 summarizes some yield data for Mollisols and Aridisols within the 254 to 365 mm (10 to 14 inch) precipitation zone from the crop-yield study.

Obviously, specific comparisons are of limited value, since widely scattered sites may have received different amounts of precipitation and different management inputs in a given year. The averages are interesting however, in 1986, winter wheat produced an average six bushels more on Mollisols than on Aridisols in the study (254 to 356 mm mean annual precipitation zone). There were only slight differences between the two groups in spring wheat or barley production. In 1987, the semi-arid Mollisols produced on the average nearly 17 bushels more spring wheat than the Aridisols, but this difference is based on only three samples for each group. There was essentially no difference between the two groups in winter wheat or barley production. In 1988, a drought year, there was insufficient data to make any comparisons. This type of crop-yield correlation data is sorely needed to assess the productivity potential of Montana's agricultural soils.

Table 1. Average dryland yields by year for Mollisols and Aridisols occurring in the 254 to 356 mm precipitation zone in Montana. Data compiled from a joint Soil Conservation Service - Montana State University crop yield study.>

Crop	Mollisols Yield Mgh^{-1}	# of obs.	Aridisols Yield Mgh^{-1}	# of obs.
1986				
Winter Wheat	2.10	30	1.69	15
Spring Wheat	2.66	15	2.49	9
Barley	2.28	12	2.21	21
1987				
Winter Wheat	2.58	27	2.60	9
Spring Wheat	3.17	3	2.04	3
Barley	2.85	4	2.77	9
1988 - Drought Year				
Winter Wheat	1.92	12	NA	0
Spring Wheat	0.87	9	NA	0
Barley	0.34	9	0.77	7

> Unpublished thesis data provided by Linda A. Spencer, graduate research assistant, Montana State University, 1989.

Table 2. SCS cropland and rangeland production estimates for selected soil series used in comparisons.>

Taxonomy/Series	Wheat>>	Wheat>>	Hay>>	Range#
	------ Mgh^{-1} ------			
Aridic Argiborolls, fine				
Scobey	2.35	2.22	3.36	1.46-1.68
Ethridge	2.35	1.75	3.36	1.46
Borollic Haplargid/Paleargid, fine				
Pinelli	2.02	1.68	2.24	1.46-1.79
Phillips	2.15	1.68	2.46	1.23-1.79
Aridic Haploboroll, fine-loamy				
Kremlin	2.35	2.15	3.36	1.46-1.79
Borollic Camborthid, fine-loamy				
Yamac	2.35	2.02	2.69	1.23-1.79
Aridic Haploboroll, fine-silty				
Floweree	2.15	1.68	2.24	1.34-1.68
Borollic Camborthid, fine-silty				
Lonna	2.29	1.75	3.36	1.23-1.79

> Data obtained from soil interpretive records (form 5's) of the Montana Soil Survey, Soil Conservation Service, USDA, 1989.
>> Estimated grain and hay yields assume a high level of management.
Potential range production for range in a good to excellent range

It does not appear likely, based on incomplete data, that surface color is an important predictor of potential dryland crop production within the semi-arid environment. This agrees with potential production estimates recorded on Soil Interpretive Records by SCS soil scientists. These estimates, based on field experience and producer averages, often show only slightly lower plant production potentials for Aridisols than for corresponding Mollisols (Table 2).

Crop yield data do not support the idea that areas receiving 254 to 356 mm of precipitation in Montana meet the criteria for an Aridic moisture regime, at least not based on the criteria of a 70% probability of crop failure in a given year.

MAPS

A decision needs to be made on the extent of the Aridic soil moisture regime in Montana. The proposed changes in the Aridisol order will alter the taxonomy of many of Montana's Aridisols in any event. Under the proposed changes, soils with mollic epipedons and an aridic soil moisture regimes will be reclassified as Aridisols. These proposed changes provide an opportunity to re-evaluate soil classification based on soil climate. The Montana Agricultural Potentials System (MAPS) developed by Caprio and

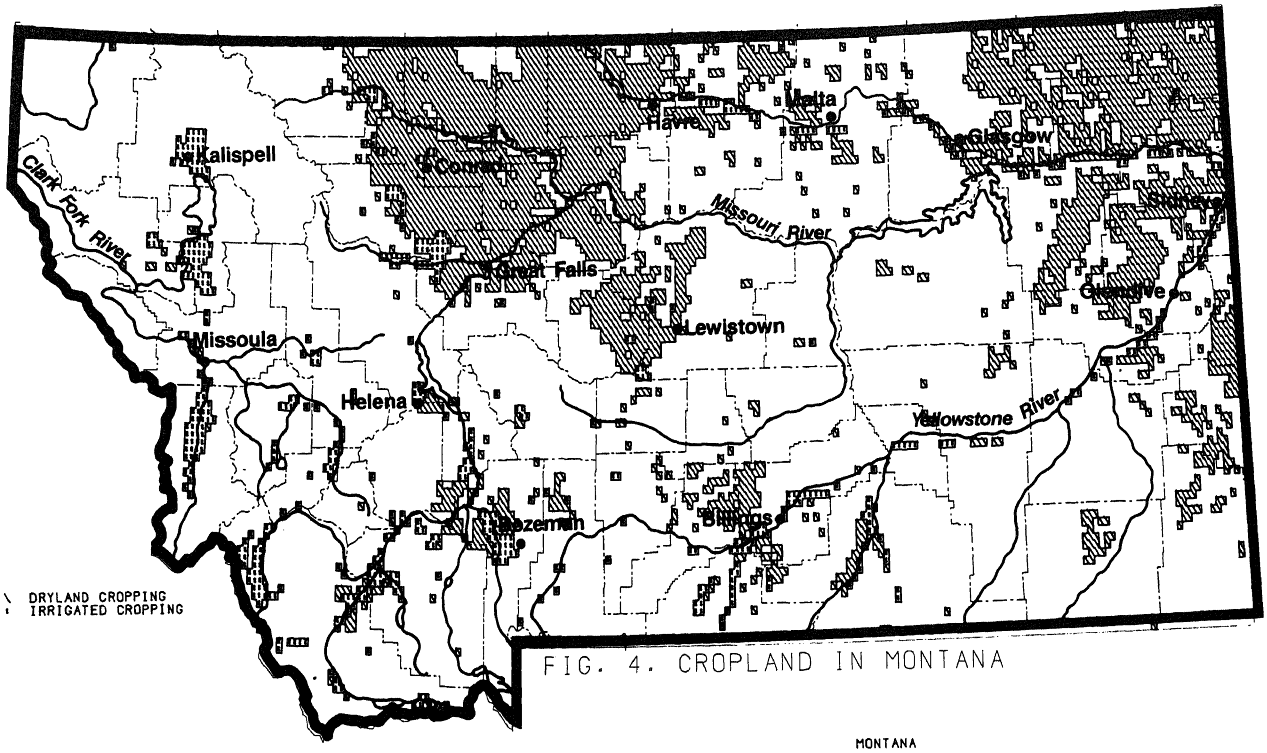


FIG. 4. CROPLAND IN MONTANA

MONTANA

ALBERS EQUAL AREA PROJECTION

PLANT & SOIL SCIENCE DEPARTMENT
MONTANA STATE UNIVERSITY, BOZEMAN, MT. ©

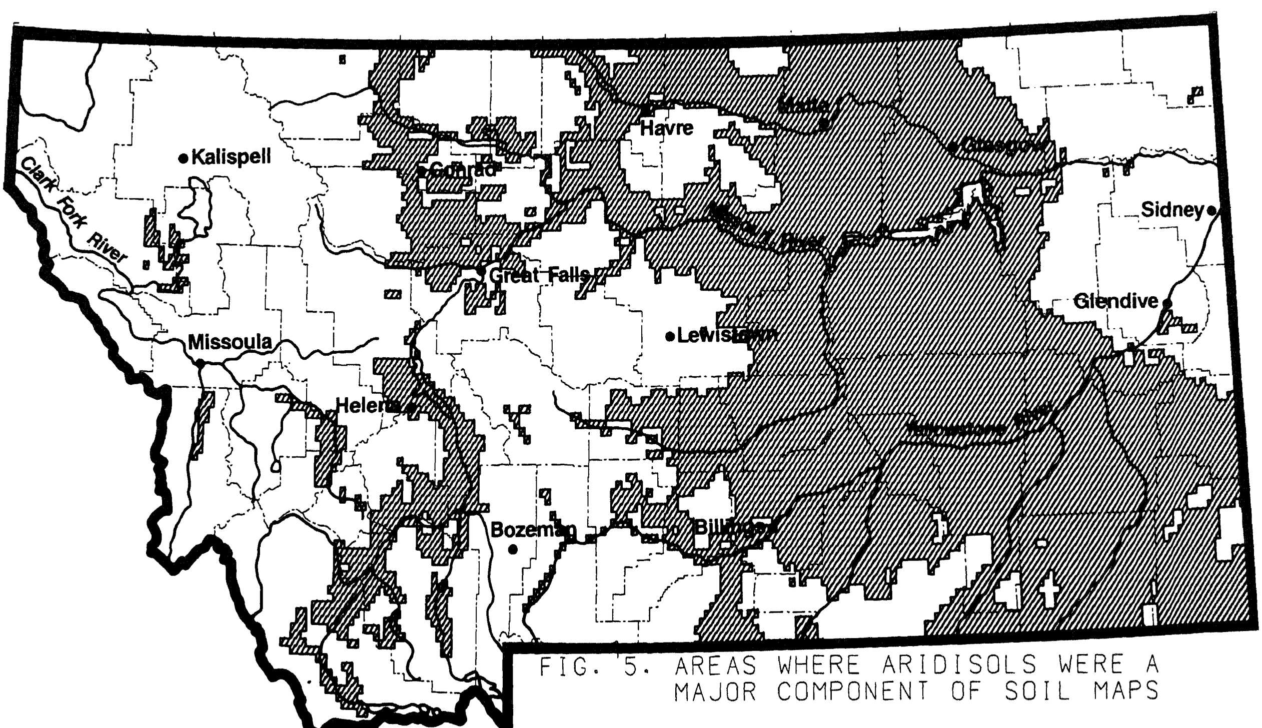


FIG. 5. AREAS WHERE ARIDISOLS WERE A MAJOR COMPONENT OF SOIL MAPS

MONTANA

ALBERS EQUAL AREA PROJECTION

Nielsen and others at Montana State University provides a valuable tool for evaluating land resources, including soil climate.

The MAPS system is a computer-driven mapping system which at this time has over 150 data layers in it. Data are stored in 18,005 cells covering the whole state. Each cell represents three minutes of latitude by three minutes of longitude or approximately eight square miles.

Fig. 1 shows the mean annual precipitation zones for the arid and semi-arid regions of Montana. Areas have been split into 305 to 356 mm (12 to 14 inch), 254 to 304 mm (10 to 12 inch), and 152 to 254 mm (6 to 10 inch) precipitation zones. Large areas of Montana receive between 254 and 356 mm of mean annual precipitation. Only two isolated areas, south of Bridger and near Dillon, receive less than 254 mm of mean annual precipitation.

Fig. 4 shows areas of irrigated and non-irrigated cropland in the state. Many of these areas, especially in the north central part of the state, correspond to areas receiving less than 356 mm of mean annual precipitation. Fig. 5 shows areas that have a large component of Aridisols mapped within them. This map is based on the general soils map for the state. It identifies all areas of the state where one or another Aridisol taxa is included in the map unit name on the general soils map. Areas not included may still have some Aridisols mapped in county soil surveys. Overlaying this map on the cropland map shows at least a general pattern in the eastern half of the state of only small amounts of cropland in predominantly Aridisol areas.

Fig. 6 shows areas where potential evapotranspiration (ET), as derived by the Thornthwaite model, exceeds measured precipitation. In general, values between 280 and 330 mm ET deficit correspond to the 254 - 304 mm precipitation zone. Values near 150 to 200 mm deficit correspond to the 305 to 356 mm precipitation zone. The dry area south of Bridger has ET deficits ranging from 356 to 432 mm. The equally low precipitation areas near Dillon have ET deficits in a range of only 254 to 304 mm. Lower temperatures in the Dillon area should be reflected in lower potential evapotranspiration, possibly indicating areas of cryic Aridisols. On the other hand, the short growing season and relatively low evapotranspiration demand in these areas may make moisture less limiting than the sparse precipitation suggests.

Fig. 7 shows areas of potential mesic soil temperature regimes based on mean annual soil

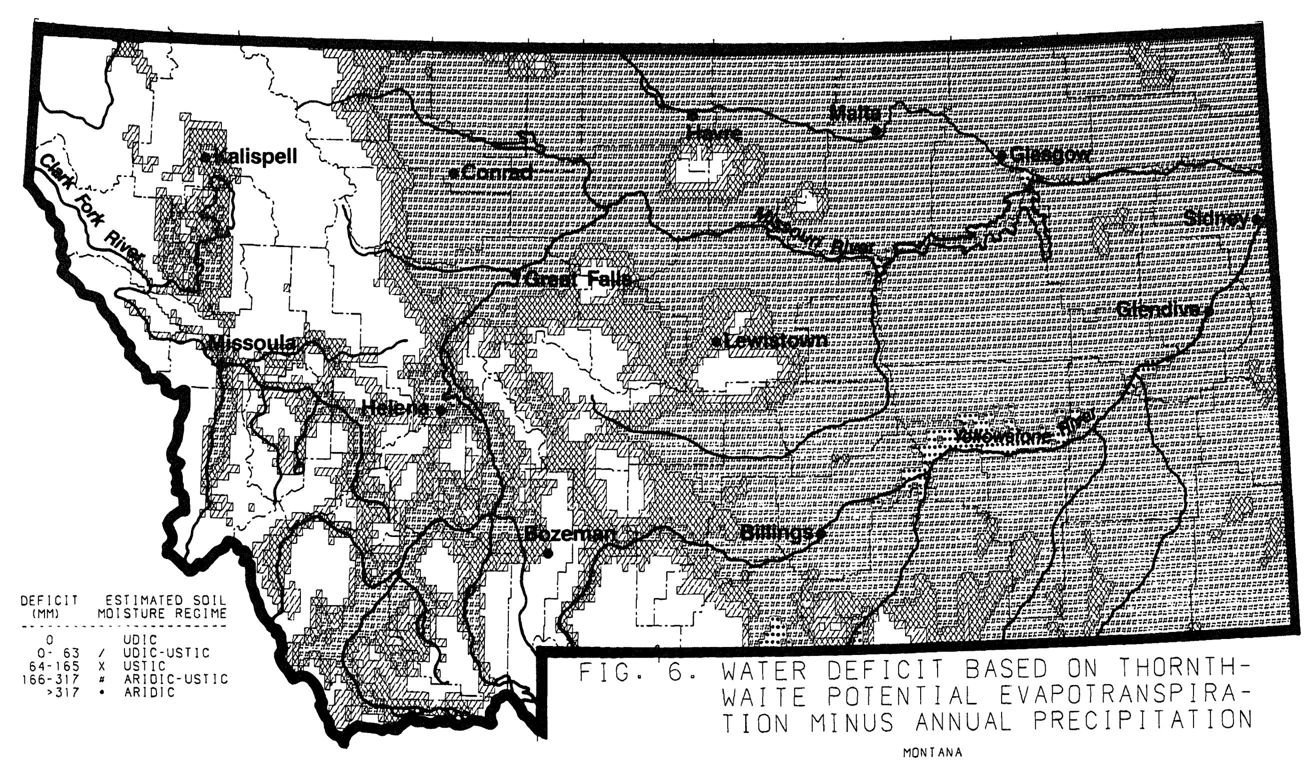


FIG. 6. WATER DEFICIT BASED ON THORNTHWAITE POTENTIAL EVAPOTRANSPIRATION MINUS ANNUAL PRECIPITATION

MONTANA

ALBERS EQUAL AREA PROJECTION

PLANT & SOIL SCIENCE DEPARTMENT
MONTANA STATE UNIVERSITY, BOZEMAN, MT. ©

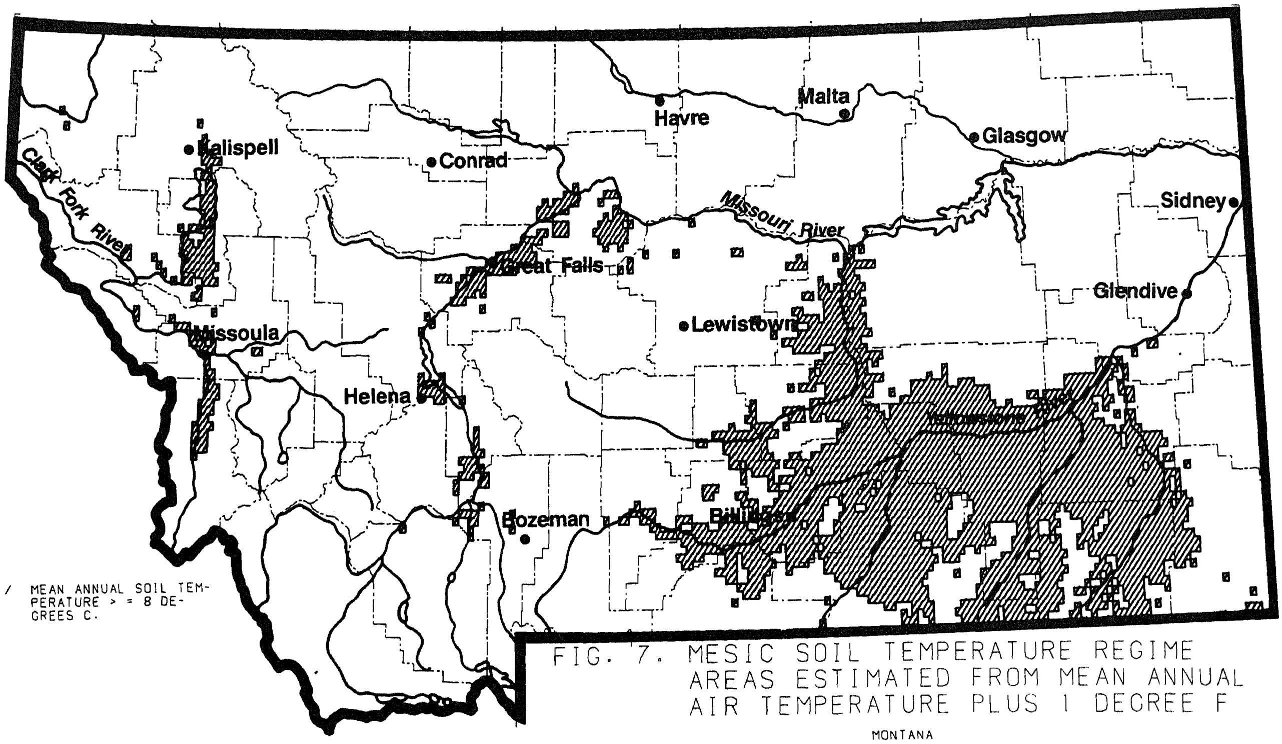


FIG. 7. MESIC SOIL TEMPERATURE REGIME AREAS ESTIMATED FROM MEAN ANNUAL AIR TEMPERATURE PLUS 1 DEGREE F

MONTANA

ALBERS EQUAL AREA PROJECTION

PLANT & SOIL SCIENCE DEPARTMENT
MONTANA STATE UNIVERSITY, BOZEMAN, MT.

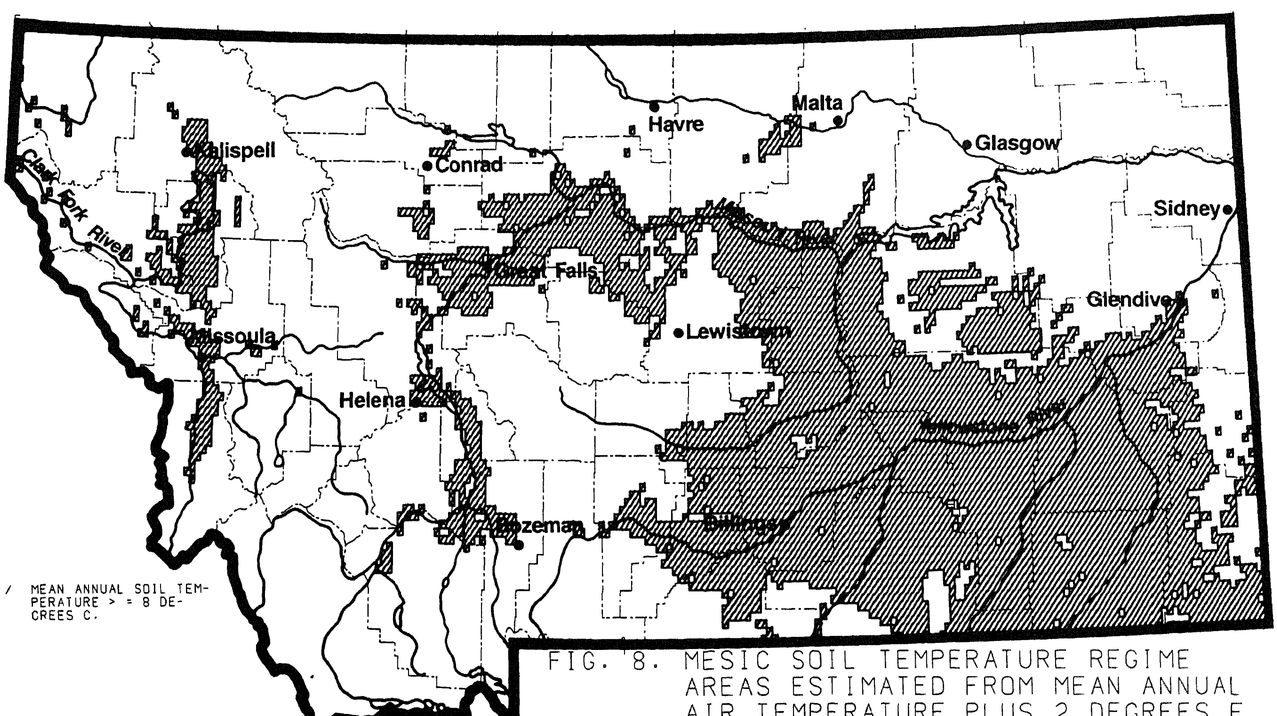


FIG. 8. MESIC SOIL TEMPERATURE REGIME AREAS ESTIMATED FROM MEAN ANNUAL AIR TEMPERATURE PLUS 2 DEGREES F

MONTANA

ALBERS EQUAL AREA PROJECTION

temperatures approximated by mean annual air temperature plus 1°F. This is a conservative estimate of mesic areas that corresponds reasonable well with areas mapped as mesic in the state or at least perceived as bordering on mesic. Fig. 8 is based upon mean annual air temperature plus 2°F and suggests a substantially larger area of mesic soil temperature regime in the state.

Soil climate data are given in Appendix 1 for each of the Montana sites on the Aridisol Tour. The data are based on the cell values in the MAPS system for these sites. Two sites, the Lonna and the Cambeth series, occur within the same MAPS cell. The Stormitt site occurs in a cell with large relief. The presence of higher elevation areas in the cell skewed the data for this site toward a wetter and cooler environment than expected for the Stormitt series. The series normally occurs in areas receiving less than 254 mm of mean annual precipitation. A separate data set was acquired from an adjacent drier cell west of the first cell.

Conclusions

Aridisols by definition must have an aridic moisture regime. They are soils that do not have water available to mesophytic plants for long periods. We cannot justifiably argue that, within the 254 to 356 mm precipitation zone, Mollisols occur within an ustic moisture regime and Aridisols occur in an aridic moisture regime. The proposed changes in Aridisol classification allow for a mollic epipedon in Aridisols; thus, a decision needs to be made regarding the soil moisture regime in these semi-arid areas.

Our recommendation at this time, is that areas receiving between 254 and 356 mm of mean annual precipitation be considered to have an ustic soil moisture regime bordering on aridic. The aridic soil moisture regime should be restricted largely to areas receiving less than 254 mm of mean annual precipitation. Many of the soils currently classified as Haplargids, Camborthids, and Calciorthids should be reclassified within the Alfisol and Inceptisol soil orders. Soils with obvious salinity or sodicity problems, which limit either available water or the infiltration of water, would still fit the criteria for an aridic soil moisture regime and be classified as Aridisols. Most of these soils do not have adequate internal drainage to be reclaimed and so the transient nature of soil salts should not be a taxonomic concern. Areas receiving less than 254 mm of mean annual precipitation represent the true aridic soil moisture regime in Montana.

These recommendations are based in part on the apparently similar organic carbon levels and yield potentials of most Aridisols and Mollisols in these semi-arid areas. The acreage of Aridisols would be greatly reduced in Montana. Many of the Camborthids and Calciorthids would be reclassified as Inceptisols. Haplargids and possibly some Paleargids would become Alfisols. The classification of Mollisols in semi-arid areas would remain unchanged.

As a final note, cryic Aridisols have not been mapped in Montana. The potential exists for using cryic Aridisols in the southwest corner of the state, but the area involved is small. If cryic Aridisols are established in Idaho, Montana could probably use them. Their establishment does not appear critical for soil mapping in Montana.

References

Aguilar, R., E.F. Kelly, and R.D. Heil. 1988. Effects of cultivation on soils in northern Great Plains rangeland. Soil Sci. Soc. Am. J. 52:1081-1085.

Anderson, D.W. 1979. Processes of humus formation and transformations in soils of the Canadian Great Plains. J. Soil Sci. 30:79-84.

Cannon, M.E. and G.A. Nielsen. 1984. Estimated productivity of range vegetation from easily measured soil characteristics. Soil Sci. Soc. Am. J. 48:1393-1397.

Decker, G.L.1972. Automatic retrieval and analysis of soil characterization data. Ph.D. Thesis. Montana State University, Bozeman, MT.

Franzmeier, D.P., G.D. Lemme, and R.J. Miles. 1985. Organic carbon in soils of North Central United States. Soil Sci. Soc. Am. Proc.49:702-708.

International Committee on Aridisols. 1989. Aridisols: Version 6.0. Soil Conservation Service, USDA, Washington, D.C. Draft, April 13.

Jenny, H. 1930. A study on the influence of climate upon the nitrogen and organic matter content of the soil. Univ. of Missouri Agric. Sta. Res. Bull. No. 152.

McDaniel, P.A., M.E. Cannon, S.J. Harvey, and G.A. Nielsen. 1982. Characterization of selected soils and vegetation on public lands in Montana. BLM Report. Plant and Soil Sci. Dept., Montana State Univ., Bozeman. 389 p.

McDaniel, P.A. and L.C. Munn. 1985. Effect of temperature on organic carbon-texture relationships in Aridisols and Mollisols.SoilSci. Soc.Am.J.49:1486-1489.

Nichols, J.D. 1984. Relation of organic carbon to soil properties and climate in the southern great plains. Soil Sci. Soc. Am. J. 48:1382-1384.

Nimlos, T.J. and D. Tippy. 1981. Soil moisture and temperature regimes in Beaverhead County, Montana. Research Note. MT For. and Sons. Exp. Sta., School of Forestry, Univ. of MT, Missoula, MT.

Shields, J.A., E.A. Paul, R.J. St. Arnaud, and W.K. Head. 1968. Spectrophotometric measurement of soil color and its relationship to moisture and organic matter. Can. J. Soil Sci.48:271-280.

Sims, Z.R. and G.A. Nielsen. 1986. Organic carbon in Montana soils as related to clay content and climate. Soil Sci.Soc.Am.J. 50:1269-1271.

Yonker, C.M., D.S. Schimel, E. Paroussis and R.D. Heil. 1988.Patterns of organic carbon accumulation in a semi-arid shortgrass steppe, Colorado.52:478-482.

Effects of Minerology and Climate on the Development of Vertic Properties in Clayey Soils of Central and Eastern Canada

C.R. de Kimpe*, C.J. Acton, C. Wang, and M.C. Nolin[1]

Abstract

Clay and heavy clay soils cover an area in excess of 1 million ha in the Lowlands of Central and Eastern Canada. These soils contain from 24 to 40% swelling minerals, the amount of which increases from the northeast to the southwest. Bulk density values, at depth, were in the range of 1.42 to 1.58 and 1.25 to 1.32 $Mg.m^{-3}$ in glaciolacustrine and marine sediments, respectively. Atterberg limits were 50 ± 6% and 26 ± 4% for the liquid and plastic limit, respectively. COLE index was generally low, but values higher than 0.15 also were recorded, indicating the potential for swell-shrink processes. A regular distribution of the precipitation throughout the year does not allow the soils to dry to a state during which slickensides may develop. The St. Lawrence Lowlands clayey soils therefore did not exhibit vertic properties. At best, some profiles could be considered as showing properties of incipient vertic development that may qualify them for vertic intergrades to other orders.

Introduction

In Central and Eastern Canada, most of the clay and heavy clay soils are located in a corridor extending from the Great Lakes northeastward within the St. Lawrence Lowland physiographic region (Bostock, 1970). Smaller areas also occur in the Canadian Shield region, especially in the Abitibi Uplands, Laurentian Highlands, around Lake St. Jean, and in the Appalachian region. From an agricultural point of view, the St. Lawrence Lowlands contains by far the most important area of clay soils for crop production and as such has prompted the interest of soil scientists in studying the characteristics and land management problems of these soils. They present various properties that are similar to the clayey soils of Western Canada, but, at the same time, striking differences do occur.

In this paper the general characteristics of Central and Eastern Canada clayey soils are summarized in relation to Vertisols and Vertic subgroups.

General Description of the St. Lawrence Lowlands

Physiography

The St. Lawrence Lowlands Region extends from the southwestern tip of Lake Erie toward the northeast, to the Strait of Belle-Isle (Bostock, 1970) (Fig. 1). It is subdivided into the Western, Central, and Eastern Lowlands. The Western Lowlands is separated from the Central Lowlands by the Frontenac axis, an extension of the Canadian Shield protruding into New York state and crossing the St. Lawrence river east of Kingston, Ontario. The Central Lowlands extends from Ottawa into the upper part of the Gulf of St. Lawrence, east of Quebec city. The Eastern Lowlands encompasses the Mingan and Anticosti islands and small areas around the Gulf and along the north-western coastline of Newfoundland.

The Lowlands cover an area of approximately 13 million ha, about 8 million ha in the western section and 5 million ha in the central and eastern sections (Matthews and Baril, 1960). Twenty-five percent of the Lowlands consists of silt and clay lacustrine and marine sediments. In Ontario, 500,000 ha were surveyed as clay and heavy clay soils, and the corresponding area for Quebec is 550,000 ha.

The major parent materials in the western section of the Lowlands have a glaciolacustrine origin. When the Wisconsinan glacier retreated, deglaciation first occurred in the Great Lakes area. Melting glaciers caused deposition in several fresh-water lakes (Dyke and Prest, 1987). Two of the larger and more recent lakes were Lake Whittlesey, a precursor of Lake Erie which formed about 13,000 years B.P., and Lake Iroquois, a precursor of Lake Ontario which formed

[1]C.R. De Kimpe and C. Wang, Agriculture Canada, LRRC, Central Experimental Farm, Ottawa, Ontario, Canada K1A 0C6; C.J. Acton, Agriculture Canada, Ontario Institute of Pedology, Guelph, Ontario, Canada, NiH 6N1; and M. Nolin, Agriculture Canada, Complexe Scientfique, 2700 Einstein, Sainte-Foy, Quebec, Canada G1P 3W8. Contribution No. 89-70, LRRC, Ottawa. *Corresponding author.

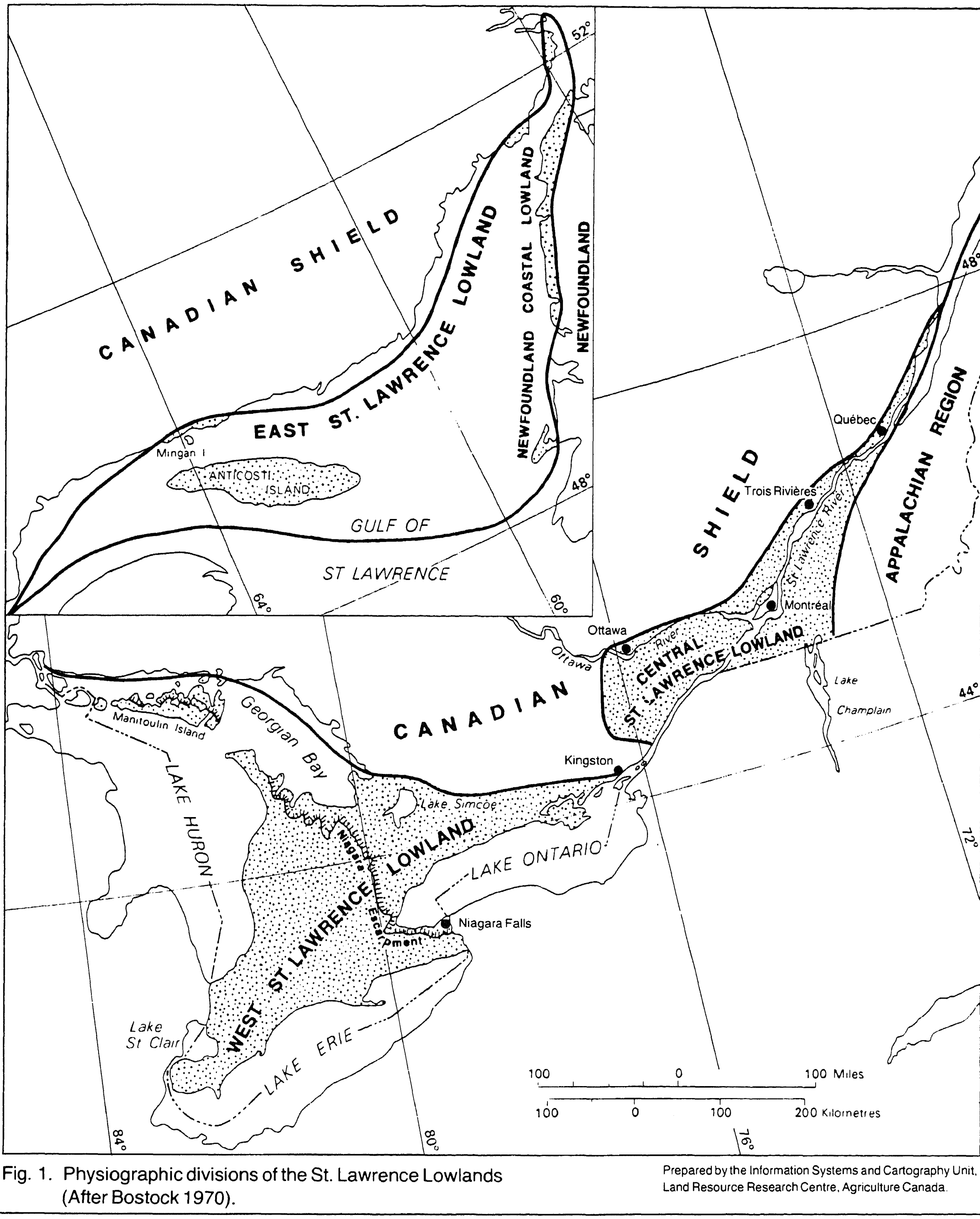


Fig. 1. Physiographic divisions of the St. Lawrence Lowlands (After Bostock 1970).

Prepared by the Information Systems and Cartography Unit, Land Resource Research Centre, Agriculture Canada.

about 12,000 years B.P. The large quantities of glacial meltwater reworked and sorted much of the till, depositing this material as lacustrine clays in these early glacial lakes. This section of the Lowlands has never been invaded by the sea, and sedimentation therefore took place in fresh water.

The Central section of the Lowlands is a large uniform plain, characterized by level to gently sloping terrain (0-3% slopes) but dissected in places by erosion by the St. Lawrence and adjacent rivers (Lajoie, 1975). The elevation within this section of the Lowlands does not exceed 40 m.a.s.l. Clay and heavy clay sediments were

deposited during the invasion and the regression of the Champlain sea, from approximately 11,800 to 9,000 years BP (Prest, 1975). The marine sediments are generally heavy clay, often calcareous with a plastic consistance. They were reworked by fluvial and water erosion and redeposited as fluviatile sediments.

The Eastern section of the Lowlands does not contain much clayey material and will therefore not be discussed further.

As most sediments in the Lowlands originate from the underlying and nearby rock formations which, over geological time, were weathered, then scoured by continental glaciers during several ice ages, it is appropriate to review briefly the bedrock geology of the Lowlands area.

Bedrock Geology

The Precambrian rocks belonging to the Grenville geological province are the easternmost extension of the Canadian Shield and form the basal bedrock formations of Southern Ontario and a part of Southern Quebec (Martini et al., 1970). These rocks outcrop north of an unconformity running from Georgian Bay to the Thousand Islands in Ontario and northeastward to Labrador along the north shore of the St. Lawrence river. Grenville rocks are strongly metamorphosed and contain metasediments (quartzites, limestones, amphibolites, paragneiss), metavolanics, and igneous rocks (anorthosites, granites, diorites, syenites, gabbros, and pegmatites).

West of the unconformity, the rock layers dip westward gently towards Michigan. Consequently, younger formations of Ordovician, Silurian, and Devonian sedimentary limestones and dolomites are found successively from east to west. Most of the rocks underlying the Lowlands in Quebec are also Ordovician dolomites, limestones, and shales, with small areas of Cambrian quartzites and sandstones.

To the south, and separated from the Canadian Shield by the Logan's fault that runs along the St. Lawrence Valley, the Appalachian Mountains consist of Paleozoic rocks, Cambrian to

Ordovician sedimentary rocks, and volcanics. The Appalachian piedmont contains limestone formations, whereas slate, schists and chloritioschists, quartzites, and sandstones are dominant on the plateaus. The rock formation of the Appalachian Mountains were weakly to moderately metamorphosed.

Climate

Temperature and Precipitation

The mildest winter temperatures are recorded near Lake Erie and St. Clair, and in the Niagara Peninsula, where January air temperature averages -5°C (Environment Canada, 1982a and b). This reflects the strong influence of the adjacent large water bodies, as considerably colder winters occur to the north and the east. January mean air temperature is -9°C at Montreal and Ottawa and -11°C at Quebec City. The mean daily temperature reaches a maximum of about 20-22°C in July throughout the Lowlands area. Based on limited data available, frost penetration in the soil does not exceed 40 cm and only for periods less than 3 months. East of Trois-Rivieres, and depending on the snow-cover, frost may penetrate deeper and stay for a longer period.

Precipitation generally is distributed uniformly during the year except in the northeastern part of the St. Lawrence Valley, where a tendency towards maximum precipitation in the summer and early autumn exists. The mean total precipitation is 960 mm at London in the southwestern most part of the Western Lowlands, decreasing to 780 mm at Toronto, and then increasing northeastward throughout the Central Lowlands. Average precipitation is 880 mm at Ottawa, 1060 mm at Montreal, and 1140 mm at Quebec city (Environment Canada, 1982a and b).

The annual potential evapotranspiration increases from 400 - 500 mm in the northeast to 500 - 600 mm west of Trois-Rivieres, Quebec, and to 600 - 700 mm in a few areas in southwestern Ontario (Clayton et al., 1977). In the May-September period, seasonal water deficit is therefore unlikely in the northeastern part of the Central Lowlands, but it occurs in the southwestern part of this section, as well as in the Western Lowlands. At the 50% probability level, for soils with 100 mm water storage capacity in the rooting zone, the deficit will be $\leq$ 200 mm, but it may be $\leq$ 300 mm if the storage capacity of the soil is reduced to 25 mm (Agriculture Canada, 1976). For 100 mm water storage capacity of the soil, the moisture index is 80-100 in the northeastern and 60-80 in the southwestern part of the Lowlands (Ibidem).

Soil Climate

The soil temperature regime is cool to moderately cool boreal in the northeast and changes to mild mesic in the area around Montreal and in

southern Ontario (Clayton et al., 1977). At some places near Lake Erie and in the Niagara Peninsula, the soil temperature regime is mild to moderately warm mesic.

Two stations will be used to illustrate the change in soil temperature over the year (Ibidem). At Fredericton, a station representing the climate of the northeastern part of the Lowlands, boreal soil climate is typically characterized, at the 50 cm depth, by a summer temperature range of 15-18°C from mid-June to mid-September and a winter temperature range of 0-2.C from early December to mid-April. Transition in the spring and the fall covers periods of about 2 months each.

At Harrow, a station in the southwestern part of the Lowlands, mesic soil climate is character ized, at the 50 cm depth, by a summer range of 18-22°C from early July to late September and a winter range of 0-2°C from early January to the end of March. In the spring, temperature increases rapidly to reach 8°C in early May, whereas in the fall it drops below that temperature by the end of November.

The soil moisture regime is perhumid in the northeast from Quebec city to Trois-Rivieres (Clayton et al. 1977), which corresponds to perudic in the *Soil Taxonomy* (Soil Survey Staff, 1975). The soil is either moist all year or seldom dry. The soil moisture regime is humid from Trois-Rivieres to Kingston and subhumid west of Kingston. These two regimes correspond to the udic moisture regime. Under the humid condition, the soil is not dry in any part as long as 90 consecutive days in most years, but there may be slight moisture deficits.

In some years under the subhumid condition, the soil is dry in some parts when soil temperature is > 5°C, which may produce significant moisture deficits. An analysis of the frequency and distribution of growing season dry spells in the years 1957-1979 indicated that short-period (10-20 day) dry spells occur at some time every year in southern Ontario (Brown and Wyllie 1984). Dry spells of longer duration (4 weeks or more) occur about once every three years.

Soils in the Lowlands

Classification

In the western section of the Lowlands, Gray Brown Luvisols (Hapludalfs) are the dominant soils on most glacial till and some lacustrine plains. Gray Brown Luvisols and Eutric Brunisols (Eutrochrepts) are commonly observed on deltaic materials and outwash plains. Luvic Gleysols (Aqualfs) and Humic Gleysols (Aquolls) are most frequently found on clayey lacustrine plains and on the poorly drained areas of the clayey textured till plains.

In the central section of the Lowlands, Eutric Brunisols and Gray Brown Luvisols are most common on till plains. Melanic Brunisols (Hapludolls) and Humic Gleysols (Aquolls) are most common on clayey marine sediments, with the Brunisols usually present on the better drained landscape position. Humo-Ferric Podzols (Haplorthods) and Dystric Brunisols (Dystrochrepts) are the dominant soils on outwash and sandy deltaic materials.

In the eastern section of the Lowlands, Humic Gleysols are most common on marine clay material, whereas Dystric Brunisols and Humo-Ferric Podzols occur more frequently on all other parent materials.

Soil Pedons

Five representative pedons will be used to illustrate the nature, properties, and general behaviour of the clay and heavy clay soils of the Lowlands.

Two pedons, of glaciolacustrine origin, were selected in the Western Lowlands. The Welland soil, sampled at an altitude of 180 m on an abandoned farmland in the Niagara region near Fort Erie, Ontario, (42° 58' N and 79° 05' W), is developed on reddish lacustrine clay derived from glacier reworking of reddish Halton clay till. It is classified as an Orthic Luvic Gleysol (Agriculture Canada Expert Committee on Soil Survey, 1987).

The Lincoln soil, sampled at an altitude of 197 m on an improved pasture field in Haldimand - Norfolk region near Cayuga, Ontario, (43° 02' N and 79° 45' W), is developed on deep-water glaciolacustrine clay. It is classified as an Orthic Humic Gleysol.

Three pedons were selected in the Central Lowlands. The Dalhousie soil, developed on marine sediments, was sampled at an altitude of 92 m on an improved pasture of the Animal Research Centre at Ottawa, Ontario, (45° 17' N and 75° 45' W). The Ste Rosalie soil, also developed on marine sediments, was sampled at an altitude of 33 m on an improved pasture at Ste Rosalie, Quebec, (45° 39' N and 72° 55' W). The Providence soil, developed on fluviatile sediments, was sampled at an altitude of 19 m on a field cropped for cereals at St. Ours, Quebec, (45° 52' N and 73° 07' W). All three soils in the Central Lowlands are Orthic Humic Gleysols.

Methods

Moist soil color was taken on the Munsell color charts. After air-drying, the soils were passed through a 2 mm-sieve. Most of the analytical procedures for the physical and chemical analyses conformed to those described in the *Manual of Methods* for soil sampling and analysis (McKeague, 1978), and they will be referred to by their code number in the manual.

- texture: pi(2.11); - bulk density: core method (2.21); - 33 kPa and 1.5 MPa moisture retention: core method (2.42) and pressure plate (2.43); - Atterberg limits: liquid (2.61) and plastic (2.62); - coefficient of linear expansion COLE: core method (2.31); - pH: in 0.01 M CaCl (3.11); - organic C: wet oxidation (3.613); - exchange capacity and exchangeable cations: in 2M NaCl (3.31); - extractable sesquioxides: in dithionite-citrate-bicarbonate (3.51) and ammonium oxalate (3.52) extracts; - surface area: EGME method (5.62).

Results and Discussion

Soil Properties

Physical properties

The soils from the central section, developed on sediments that had been, at least at some time, under marine influence, had a 5Y hue in the deeper horizons (Table 1). Soils from the western section, developed on glaciolacustrine sediments, were influenced by the nature of the parent material, resulting in hues of 10 YR in the Lincoln soil and 7.5 YR in the Welland soil. Organic matter gave a slightly redder hue, generally 7.5 YR, in the surface horizon of most soils.

All soils had a clay to heavy clay texture at depth. However, the Providence soil profile also had a Cg horizon with silty clay texture. This was related to the fluviatile origin of the parent material, which may be interspersed with coarser material. In the solum, texture was somewhat coarser than at depth, ranging from silty clay loam to clay, but only the Welland pedon showed evidence of clay illuviation (Table 1).

Bulk density (Table 1) of the deep horizons was greater in the glaciolacustrine (1.42 - 1.58) than in the marine sediments (1.25 - 1.32 Mg m^{-3}). Values in the same range were recorded in several soils developed on marine sediments in Quebec (De Kimpe and Mehuys, 1979). The difference in bulk density between the two types of parent materials probably resulted from the sedimentation in quiet versus turbulent waters.

Larger values for the water retention at 33 kPa and 1.5 MPa were determined for the soils from southern Ontario and Ottawa (>40 and 25-35%) than for those located east of Montreal (30-40% and 20-35%) (Table 1). The difference is thought to be related to the different content of swelling minerals and fine clay in the soils of the two areas.

Atterberg limits were comparable for all soils, with values of 50 $\pm$ 6% for the liquid limit and 26 $\pm$ 4% for the plastic limit (Table 1). It has been determined, for a large number of soils with a wide range in texture and organic matter content, that Atterberg limits were related more to organic matter than to the clay content (De Kimpe et al., 1982). However, this conclusion was not verified in the clay and heavy clay soils of this study.

The K_{sat} values were closely related to soil structure and farming practices (Wang et al., 1985). The soils developed on marine clay, Dalhousie and Ste Rosalie, had strongly developed blocky structure and numerous biopores in the subsoil (Table 1). Therefore, the corresponding K sat values at depths > 50 cm were high. The other three clayey soils developed on glaciofluviatile, glaciolacustrine material and till and had weakly or moderately expressed soil structure and few biopores and, in the case of the Welland and Lincoln soils, also a high bulk density. These soils had low to very low K_{sat} values in the subsoil horizons. The surface Ap horizons generally had a much greater K_{sat} value than the underlying B horizon because of the frequent plowing and lower bulk density.

The COLE parameter was generally low in all soils of the Lowlands (Table 1). The lowest values, $\leq$ 0.05, were recorded for the Ste Rosalie and Providence soils, although a value of 0.15 was recorded in a Ste Rosalie soil profile (R. Asselin, personal communication, 1989). The COLE parameter was higher, 0.15, in the Dalhousie soil than in the Ste. Rosalie and Providence profiles. In the Welland and Lincoln soils from southwestern Ontario, values of about 0.10 and slightly higher values, respectively, were obtained. This range of COLE values was likely related to the presence and the nature of the swelling minerals and to the high proportion of fine clay.

Table 1. Selected physical properties of the five pedons.

Horizon	Depth (cm)	Color	Structure=	Clay (%)	Silt (%)	Bulk density (Mg/m^3)	Moisture retention 33 kPa (%)	Moisture retention 1.5 MPa (%)	Atterberg limits liquid (%)	Atterberg limits plastic (%)	COLE index	K_{sat} (cm/h)
Ste Rosalie (marine origin) Aquoll												
Ap	0-28	10YR 3/1	str.f. sub. bl.	43	49	1.26	33	21	43	25	0.04	12
Bg1	28-42	2.5 Y 5/1	wk.f. sub. bl.	53	44	1.41	34	25	47	21	0.04	7
Bg2	42-56	5Y 5/1.5	mod.f. sub. bl.	65	35	1.35	40	30	55	24	0.05	23
Bg3	56-100	5Y 5/2	mod.f. sub. bl.	72	28	1.32	45	31	55	25	0.05	29
Cg	100-140	5Y 5/1.5	str.f. to med.a.bl.	72	27	n.d. +	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ckg	140+	5Y 6/1	str.med. to c.a.bl.	68	32	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Dalhousie (marine origin) Aquoll												
Ap	0-17	10YR 2/2	str.f. sub. bl.	43	41	n.d.	44	18	56	40	0.11	15
Bg1	17-24	5GY 6/1	wk.f. sub. bl.	31	37	1.37	37	15	n.d.	n.d.	0.09	6
Bg2	24-53	5Y 4/2	mod.med. sub. bl.	62	34	1.37	43	29	45	27	0.15	12
BCg	53-75	5Y 4.5/1.5	mod.med. sub. bl.	65	34	n.d.	40	27	n.d.	n.d.	0.15	20
Cg	75-100	5Y 4/1.5	str.med. to c. bl.	65	34	1.26	47	26	47	23	0.15	35
Providence (fluviatile origin) Aquoll												
Ap	0-27	2.5 Y 3.5/2	massive	37	52	1.30	38	23	44	28	0.04	58
Bg1	27-44	10YR 5.5/1	massive	44	49	1.35	41	25	44	25	0.05	2
Bg2	44-70	2.5Y 5/3	wk.f. sub. bl.	52	45	1.33	43	26	50	25	0.05	<1
Bg3	70-94	2.5Y 5.5/3	wk.med. sub. bl.	58	39	1.27	47	28	54	25	0.05	<1
Cg1	94-122	5Y 5/2	mod.c. sub. bl.	59	37	1.25	n.d.	n.d.	55	24	n.d.	<1
Cg2	122-170	5Y 5.5/2	n.d.	47	50	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	<1
Welland (reworked till) Aqualf												
Ap	0-15	10YR 4/2	mod.c. gran.	48	45	1.16	38	20	46	28	0.07	9
Btg1	15-34	7.5YR 4/2	mod.c. sub. bl.	69	28	1.40	51	31	62	26	0.09	<1
Btg2	34-43	7.5YR 5/2	mod.c. col.	77	22	1.42	52	36	n.d.	n.d.	0.10	n.d.
Ckg	43+	7.5YR 5/2	wk.c. col.	73	26	1.42	48	35	56	28	0.08	<1
Lincoln (lacustrine origin) Aquoll												
Ap	0-15	10YR 3.5/3	mod.c. bl.	48	47	1.28	43	27	43	27	0.10	33
Bgf	15-40	10YR 4/1	wk.c. bl.	62	34	1.49	36	25	60	26	0.20	<1
BCg	40-50	10YR 4/1	wk.c. bl.	57	37	n.d.	n.d.	n.d.	44	22	0.10	n.d.
Ckg	50+	10YR 5/2	mod.c. bl.	60	38	1.58	33	21	51	22	n.d.	<1

+ n.d. = not determined
= mod. = moderate, str. = strong, wk. = weak, f. = fine, med. = medium, c. = coarse, sub. = subangular, bl. = blocky, col. = columnar.

Chemical properties

All soil profiles had pH values ≥ 6.8 in the deeper horizons (Table 2). In the Welland and Lincoln soils, values > 7.4 at depth were related to the presence of carbonates in the Paleozoic rock formations from which the sediments were derived. In the Ste Rosalie and Providence soils, pH values of 7.5 were attributed to the presence of fossil shells in these marine sediments.

Accumulations of organic carbon were mostly restricted to the surface horizon, as is commonly observed in poorly drained soil pedons (De Kimpe et al., 1979). Content ranged from 1.9 to 9.2% in the Ap horizon, whereas it decreased to 0.4 to 1% in the Bg horizons.

Cation exchange capacity was generally < 30 $cmol/kg^{-1}$. The high value for the Dalhousie Ap horizon was related to its high organic matter content. The most abundant cation on the exchange sites was Ca^{2+} , followed by Mg^{2+}.

Dithionite-citrate-bicarbonate extractable Fe content ranged from 0.5 to 3.1%, with the highest values being found in the Southern Ontario soils. This reflected in the soil color (Table 1). In the Welland and Providence soil pedons, the DCB extractable Fe content decreased with depth, whereas there was a maximum at the Bg

Table 2. Selected chemical properties of the profiles.

Horizon	pH (in 0.01 M $CaCl_2$)	Org C (%)	CEC (cmol/kg)	Fe_d (%)	Fe_o (%)	Al_d (%)	Al_o (%)
Ste Rosalie							
Ap	6.2	2.3	27	0.7	0.4	0.2	0.2
Bg1	6.5	0.4	27	0.9	0.4	0.2	0.2
Bg2	6.8	0.3	28	0.9	0.5	0.3	0.3
Bg3	7.0	0.3	25	0.9	0.5	0.2	0.3
Cg	7.5	0.3	25	0.9	0.5	0.2	0.3
Ckg	7.5	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Dalhousie							
Ap	6.6	9.2	56	0.6	n.d.	0.2	n.d.
Bg1	6.9	1.0	23	0.5	n.d.	0.1	n.d.
Bg2	6.9	0.6	34	0.9	n.d.	0.2	n.d.
BCg	6.9	0.4	27	0.7	n.d.	0.2	n.d.
Cg	6.8	0.2	27	0.6	n.d.	0.2	n.d.
Providence							
Ap	6.0	1.9	28	1.6	1.1	0.3	0.4
Bg1	5.9	0.9	27	1.3	0.9	0.3	0.4
Bg2	6.5	0.2	27	1.0	0.4	0.3	0.3
Bg3	7.1	0.1	28	1.0	0.5	0.3	0.4
Cg1	7.3	0.1	27	1.0	0.5	0.3	0.4
Cg2	7.5	0.1	23	1.0	0.4	0.3	0.3
Welland							
Ap	5.2	2.4	17	1.7	0.5	0.2	0.2
Btg1	6.5	0.6	29	1.7	0.3	0.2	0.2
Btg2	7.3	0.6	31	1.5	0.2	0.2	0.2
Ckg	7.7	n.d.	29	1.2	0.2	0.1	0.2
Lincoln							
Ap	6.4	3.3	n.d.	1.7	0.5	0.3	0.2
Bgf	6.0	0.9	27	3.1	0.7	0.3	0.2
BCg	7.1	0.5	n.d.	1.7	0.5	0.2	0.2
Cjg	7.9	0.5	25	1.3	0.2	0.1	0.1

horizon level in the three other pedons. The oxalate-extractable Fe content was <1.1%. The percentage of Fe present as crystalline Fe oxides (Fe_{DCB} -Fe_o) ranged from 0.4 to 2.4%. The amount was highest in the Bg horizons of all pedons and also 2 to 4 times larger in the soils from southern Ontario than in those from the Ottawa-St. Lawrence Lowlands.

Extractable Al content was <0.4% in all soils and generally similar in both extracts.

Mineralogical properties

In the northeastern part of the Central Lowlands, illite and chlorite were the dominant phyllosilicates in the clay fraction and their sum amounted to about 40% throughout the profiles (De Kimpe et al., 1979). Smectite and vermiculite contents were about 15 and 9%, respectively. In the Montreal area of the Central Lowlands, illite plus chlorite were the major phyllosilicates with an average sum content of about 23 and 30% at the surface and at depth, respectively. Smectite content was about 18% at the surface and 16% at depth, whereas vermiculite was about 11% throughout the pedons (Ibidem).

In both regions, quartz, about 20%, feldspar, with 5-10%, and other primary minerals such as hornblende, present in minor to low amounts in the clay-size fraction, reflected the origin of the sediments and underlined the effect of the continental glaciers that produced rock flour when they scoured the rock formations. The persistance of weatherable minerals and the small changes in the mineralogical composition demonstrated the low to moderate rate of weathering in these cool to moderately cool and wet soils (Kodama, 1979; De Kimpe and Martel, 1986).

In the Dalhousie soil of the Central Lowlands, smectite content of the clay fraction increased to 35% at the surface from 22% at depth, whereas the micas decreased to 8% at the surface from 18% at depth indicating pedogenic weathering of mica to smectite (Ross et al., 1987).

Swelling minerals and illite were the major phyllosilicates in the Lincoln soil (unpublished data) of the Western Lowlands, with the swelling minerals dominant in the Ap horizon. In the lower horizons, illite was associated with small to trace amounts of chlorite, quartz and feldspar. A similar composition was found in the Welland soil (unpublished data). Swelling minerals and illite were dominant throughout the pedons and some chlorite was present in the BC horizon. Feldspar and quartz were also present in small amounts in the clay fraction of this soil.

Specific surface area was determined for clay soils from the central and western sections of the Lowlands. Average values of 245 and 210 m^2g^{-1} were obtained for the clay fraction of Gleysolic soils in the Montreal and Quebec areas, respectively (De Kimpe et al., 1979). Multiple regression analysis of 34 samples of heavy textured Humic Gleysols from southwestern Ontario produced a value of 207 m^2g^{-1} for the clay fraction (Evans, 1982). These figures supported the mineralogical composition obtained from X-ray diffraction.

Potential for Vertic Properties Development

Ahmad (1983) reviewed the literature on Vertisols and summarized their outstanding features. Although these soils are found under a wide range of climatic conditions, one common characteristic is the seasonality of the precipitation allowing for annual wetting and drying of the solum. These soils have a clay content of ≥ 35%, most often montmorillonitic, that causes pronounced changes in volume with changes in water content, resulting in deep, wide cracks in the dry seasons. The soils are plastic and have a sticky consistency when wet. Bulk density is high in dry soils and K sat is low in wet soils.

There is an appreciable increase in microelevation of the soil surface as the soil becomes wet, and, when it dries, subsidence occurs and cracks develop. In most cases, infilling of the cracks with surface material creates pedoturbation and tends to restrict strong profile development. Internal stresses due to overburden pressure, as well as shrinking and swelling of the subsoil, are at the origin of slickenside formation, and a gilgai microrelief is commonly observed at the soil surface. Surface cracking and slickensides are the major features taken into consideration for the classification of Vertisols (Soil Survey Staff, 1975; Soil Survey Staff, 1987).

The clay and heavy clay soils of the St. Lawrence Lowlands of Central and Eastern Canada, based on their texture, were well within the range of the Vertisols. From the mineralogical point of view, although they were not dominantly montmorillonitic, these soils contained significant amounts of swelling minerals which are important to the shrink-swell process. They also contained illite that, under adequate weathering conditions, transforms to swelling minerals.

All soils but Welland had a strong to moderate blocky structure at depth, which was inher-

ited from the parent material. Welland had weak columnar structure at depth. In the B horizons, the structure was affected by pedogenesis. In the three pedons from the Central section, soil structure became finer and less strongly expressed than in the C horizons, whereas coarse blocky structure was common in the B horizons of the two pedons from the Western Lowlands. Blocky structure is conducive to the development of vertic properties (McCormack and Wilding, 1973) through the wetting and drying cycle of the peds.

However, the limiting factor for the development of vertic properties in this region of Canada is soil climate. As detailed in an earlier section, precipitation is rather uniformly distributed throughout the year, but with limited probabilities for dry spells. Under the perhumid soil moisture regime prevailing in the northeastern part of the Central section, cracking patterns developed only during exceptionally dry summers, such as in 1988, and the cracks did not penetrate very deeply in the soil.

The humid soil moisture regime in the Ottawa-Montreal area is characterized by dry spells of slightly longer duration so that cracking may occur more frequently, but again not to a sufficient degree to develop permanent vertic features. Vertical soil displacement of 10% throughout the year has been monitored in a Rideau clay soil (K. Wires, personal communication, 1989). The cracking pattern was also particularly well developed during the dry summer of 1988.

The longest dry spells are regularly observed in the subhumid western section of the Lowlands. Cracking patterns develop every year and in 1988, the cracks developed vertically to a depth of 75-90 cm (K. Wires, personal communication, 1989). It is not clear, however, whether the cracks are permanent from year to year or whether sufficiently strong internal stresses occur to induce development of slickensides. The permanent development of such vertic features has not been reported to date in the clayey soils of Eastern Canada.

Literature Cited

Agriculture Canada. 976. Agroclimatic atlas. Agrometeorology Research and Service Section, CBRI, Ottawa, Ontario, 17 maps.

Agriculture Canada Expert Committee on Soil Survey. 1987. The Canadian system of soil classification. 2nd edition. Agriculture Canada Publ.1646, Ottawa, 164 p.

Ahmad, N. 1983. Vertisols. p.91-123 In: L.P. Wilding, N.E. Smeck and G.F. Hall (eds). Pedogenesis and soil taxonomy. Vol II. The Soil orders. Elsevier Publ., New York.

Bostock, H.S. 1970. Physiographic subdivisions of Canada. p.10-30 In: R.J.W. Douglas (ed) Geologic and Economic Minerals of Canada. Geological Survey of Canada. Economic Geology Report no. 1, Ottawa.

Brown, D.M. and W.D. Wyllie. 1984. Growing season dry spells in southern Ontario. Climatology Bull. 18:15-30.

Clayton, J.S., W.A. Ehrlich, D.B. Cann, J.H. Day, and I.B. Marshall. 1977. Soils of Canada. Vol. I. Soil Report. Can. Dept. of Agriculture, Ottawa, 243 p.

De Kimpe, C.R., M. Bernier-Cardou and P. Jolicoeur. 1982. Compaction and swelling of Quebec soils in relation to their soil-water properties. Can. J. Soil Sci. 62:165-175.

De Kimpe, C.R., M.R. Laverdi re and Y.A. Martel. 1979. Surface area and exchange capacity of clay in relation to the mineralogical composition of gleysolic soils. Can. J. Soil Sci. 59:341-347.

De Kimpe, C.R. and Y.A. Martel. 1986. Retrospective des recherches sur la genese et la classification des sols au Quebec. Cahiers de l'ACFAS no 37, 88-114.

De Kimpe, C.R. and G.R. Mehuys. 1979. Physical properties of gleysolic soils in the lowlands of Quebec. Can. J. Soil Sci. 59:69-78.

Dyke, A.S. and Prest, V.K. 1987. The late Wisconsinan and Holocene history of the Laurentide ice sheet. Geographie Physique et Quaternaire. 61:237-263.

Environment Canada. 1982a. Canadian climate normals 1951-1980. Temperature and Precipitation Ontario. Canadian Climate Program, 254 p.

Environnement Canada. 1982b. Canadian climate normals 1951-1980. Temperature and Precipitation Quebec. Canadian Climate Program, 216 p.

Evans, L.J. 1982. Cation exchange capacities and surface areas of Humic Gleysolic Ap horizons from southwestern Ontario. Can. J. Soil Sci. 62:291-296.

Kodama, H. 1979. Clay Minerals in Canadian Soils: their origin, distribution and alteration. Can. J. Soil Sci. 59:37-58.

Lajoie, P. 1975. Agricultural lands in southern Quebec: distribution, extent, and quality. Agriculture Canada Publ. 1556, Ottawa, 62 p.

Martini, I.P., R. Protz and W. Chesworth. 1970. The rocks and soils of southern Ontario. Dept. of Soil Science, Univ. of Guelph, 11 p.

Matthews, B.C. and R.W. Baril. 1960. The soils of the Great Lakes - St. Lawrence Lowlands. Agricultural Institute Review, 37-40.

McCormack, D.E. and L.P. Wilding. 1973. Proposed origin of lattisepic fabric. Pages 761-771. In G.K. Rutherford (ed) Soil Microscopy 4th Int. Working Meeting on Soil Micromorphology. The Limestone Press, Kingston, Ontario.

McKeague, J.A. (ed). 1978. Manual of methods for soil sampling Office, Washington D.C.

Soil Survey Staff. 1987. Keys to Soil Taxonomy (third printing) SMSS technical monography no 6, Ithaca, New York.

Wang, C., J.A. McKeague, and K.D. Switzer-Howse. 1985. Saturated hydraulic conductivity as an indicator of structural degradation in clayey soils of Ottawa area, Canada. Soil and Tillage Res. 5:19-31.

Procedures and Rationale for Development of Adequate and Comprehensive Field Soil Data Bases

R. Langohr[1]

Abstract

In most pedon data bases, data are missing for an unequivocal classification of the soils or for an exact land evaluation of the soil site. The paper describes the procedures and rationale followed by a team of pedologists of the University of Ghent in order to elaborate new guidelines for the development of more adequate and comprehensive field soil data bases. The surveyor's attention is particularly asked for features such as, producing more numerical data, to draw detailed figures of both vertical and horizontal sections, to inform about errors, temporal variability, activity status, spatial variability, related distributions and to provide his opinion about the genesis of the characteristics he describes. The final goal of the project is to raise the quality of the field data base to a level where it would permit to understand better the soil-ecosystem dynamics and where it would allow more applications than it is today.

Introduction

A series of research projects performed in collaboration with M.Sc and Ph.D fellows at the International Training Centre for Post Graduate Soil Scientists at the University of Ghent has allowed intensive evaluation of soil data bases for soil classification and land evaluation. Published soil pedon data including soil site, soil profile and soil analytical data (the whole set of these data is called here the "pedon data base" or PDB) were rigorously applied to soil classification (mainly the USDA and the FAO-Unesco soil taxonomies) and land evaluation diagnostic criteria (Adiwiganda, 1986; Lopulisa, 1986; Lopulisa et al., 1986a and 1986b; Estoista, 1988; Vaca, 1988). These investigations showed that in most PDBs, data are missing for an unequivocal classification of the soil or for an exact land evaluation of the soil site.

Similar problems, but of a larger magnitude, are met when PDBs are consulted for interpretations and for advise on plant and faunal ecology, ecosystem dynamics and environmental management. Furthermore, almost all PDBs are particularly inadequate in terms of field data needed for soil genesis investigations and paleoenvironmental reconstruction.

As the soil is largely that part of the earth where the biosphere and the geosphere meet and interact, it represents an essential source of information for many disciplines. That is why the soil surveyors, among all scientists working in the field, should have knowledge of the largest spectrum of disciplines; soil survey reports thus can contribute data for studies in ecology, environment, geomorphology, Quaternary geology, archaeology etc., if the reports are sufficiently comprehensive and adequate.

The above mentioned criticisms can be summarized in two statements.

1. The standard guidelines for profile description and analysis are written in the framework of traditional soil cartography and associated soil classification system(s). If the existing guidelines are followed rigorously by the surveyors and laboratory staff the PDBs would be adequate for these purposes; however, they are not and therefore professiona qualification of the scientists can be questioned here.
2. When it comes to other types of research projects, the standard guidelines do not provide all the required data. Yet in many projects these same standard guidelines are followed with as a result numerous unsatisfied customers of the PDBs. Therefore new guidelines should be developed and completed with special training of the surveyors.

Concerning the second statement the group of the University of Ghent is trying to improve the method of soil data recording in the field.

Our five main goals are:

1. To make the field pedon data base (FPDB) more adequate for applications known at present,
2. To make the FPDB more comprehensive in a way that it may serve in the future as a basic source of information for more disciplines than those for which it is used today,

[1]Department of General Pedology, University of Gnent, Krijgslaan 281, B-900-Gnent, Belgium.

3. To raise the quality of the FPDB to a level at least equivalent to the present-day soils laboratory data bases,
4. To raise the quality of the FPDB to a level where the computer can be used more efficiently, instead of lowering its quality in order to make it more computer-compatible
5. To make the FPDB more accessible to non-pedologists.

Summary of Procedures and Rationale for Field Guidelines

This paper describes guidelines developed by University of Ghent investigators for the purpose of raising the quality, comprehensiveness and adequacy of FPDBs. The term *standard guidelines* used here refers to those guidelines for soil profile description elaborated for sustaining regional or national soil survey activities (e.g. Soil Survey Staff 1951, Hodgson 1976, FAO 1977, FitzPatrick 1977).

Parent material, groundwater table, soil color, structure, etc. are called *soil characteristics*. Further distinctions of these characteristics are called "aspects" of soil characteristics, such as 1) origin-, age-, texture-, and mineralogical composition- of parent material, 2) origin-, degree of stagnation-, chemical composition- and color- of groundwater, 3) moist-, dry-, 4) rubbed soil color and 5) size-, shape- and grade for soil structure.

As there are numerous different types of applications of PDBs, the surveyor should **select** from the guidelines those characteristics which are relevant for his investigation(s). It must be clear that the goal of the guidelines is that not all soil characteristics and aspects of the characteristics should be described systematically.

No special effort should be made to avoid *overlapping of information*. Data about a particular characteristic may be asked for in several paragraphs of the description; if it is relevant in the study, information should be provided each time, or reference should be made to the paragraph where the information is provided in sufficient detail.

Exact and detailed *references* should be mentioned for all the data extracted from publications, reports, maps, photographs, and figures. By noting the existence of this information the user of the PDB will be able to consult these documents whenever more detailed information is needed.

Pre-established classes such as "few-, common-, abundant-", "fine-, medium-, coarse-", "abrupt-, clear-, gradual-, diffuse-", should be avoided as much as possible and should be replaced by direct measurements or estimates expressed as *numerical data*. At a time when more and more data are stored in computers, it is illogical to force the field surveyor to provide part of his data in pre-established classes; this represents a serious loss of precision and in fact amputates the use of a computer to calculate classes for us. It is still possible to derive "class" data from numerical data whenever that is useful or necessary.

Whenever a feature is estimated or measured, an error will be made. In pedology, most guidelines for soil description, and even the handbooks describing laboratory methods, focus little attention on *absolute and relative errors* associated with estimates and/or measurements. The absence of this information reduces the data's value.

Special attention should be given to records of the temporal variability (= TV), the activity status (= AT) and the spatial variability (= SV) of many of the characteristics.

Some soil characteristics like surface crusting-, cracking pattern-, mottling- and grade of aggregate development- may vary seasonally. Other characteristics such as erosion- and inundation- may have variability over longer time spans. Aspects of these characteristics such as "duration-" and "frequency-" provide information about *temporal variability (= TV)*.

The concept of *Activity Status (= AS)* is to ascertain the degree of temporal variability; here the soil scientist would note if the observed characteristic is the result of a still active process or not; in latter conditions the characteristic (e.g. microrelief-, gullies-, mottling-, biogalleries-) will be considered to be a relict, probably associated to past and different environmental conditions.

Many recent papers focus attention on *spatial variability (= SV)* of soil characteristics. Handbooks for recording soil data guide surveyors in the collection of spatial variability data for site descriptions as well as for profile descriptions. For example, information about vegetation-, erosion-, and microrelief should be described separately for the spot of the profile itself and for the surroundings. In many profile descriptions it is not clear what information belongs to the

exact location of the profile, and what information concerns the surroundings or even the entire mapping unit; these data are equally important but the spatial position should be specified and specific.

As for variability *within the profile*, it is interesting to quote Soil Taxonomy (Soil Survey Staff, 1975, p. 3) :

> "A pedon has the smallest area for which we should describe and sample the soil to represent the nature and arrangement of its horizons and variability in the other properties that are preserved in the samples...Its lateral dimensions are large enough to represent the nature of any horizons and variability that may be present...The area of a pedon ranges from 1 to 10 square metres, depending on the variability of the soil."

It is our experience that soil surveyors very seldom check this variability of the soil characteristics over an area of 10 square metres, or even one square metre. Consequently we know little about this aspect of the soil characteristics. Guidelines should systematically ask for this information and not simply mention it in an introductory chapter.

Much more effort should go to *the explanation* of the site and profile characteristics beyond the usual profile description. This part of the data base can eventually be written in a way that also non-professionals can read and understand the text. Important information about the "origin" of site and soil characteristics can be obtained from the field surveyors. It is an enormous loss if surveyors with their huge amount of unique professional skills are not asked to record their interpretations, thoughts and doubts about the genesis, the origin and the dynamics of the features described. For many characteristics, laboratory data will not provide much of this type of information. In a recent call for more Field Soil Science R.B. Daniels (1988) wrote :"A soil sample in the laboratory is nothing more than a bag of dirt. That bag of dirt becomes a useful research sample only if we know the field relations it represents".

Inclusion of these extra data focuses the surveyor's attention on additional relevant field data. If answers are needed from laboratory data, sampling is also focused and better designed to obtain exactly those laboratory data that provide the best or most usable information.

Particular emphasis should also go to the detection, description and explanation of *related distributions (= RD)* (e.g. silt capping on stones, pseudo-gley along pores, position of clay coatings-, mangans-, ferrans-, secondary carbonates-...). Except for a few characteristics like coatings, these data are largely absent from most handbooks for soil description. It is evident that related distributions, if present, should be explained.

In describing soil profiles, as much attention should go to the observation and description of *horizontal sections* as to the traditional vertical sections. Some characteristics such as cracking-, earthworm galleries-, root density-, mottling pattern- can only be described and quantified correctly by the observation of horizontal sections. Numerous studies of horizontal sections of soil profiles make us conclude that by adding this type of observation the surveyor obtains a much more complete and often even a different picture of the soil than by only observing the vertical profile walls.

Numerous *figures* should be part of the FSDB. Many characteristics of the soil and its environment can be described best with the help of figures. Topographical location-, lithostratigraphy of the parent material and of the substratum-, geomorphology-, microrelief-, associated soils-, vegetation-, soil surface form-, soil horizon distribution in the vertical profile section-, distribution patterns of particular soil characteristics such as galleries-, mottling-, carbonate accumulations- in the horizontal as well as in the vertical profile sections, are all examples of soil characteristics for which the description can be drastically improved and yet at the same time considerably simplified by adding good quality figures. Text and words should be complementary to the figures rather than the reverse. Figures are of enormous help in the description of the sometimes complicated spatial variability and related distributions of some soil characteristics. They can show the exact location of sampling sites. Furthermore, drawing figures forces the pedologist to observe very precisely a series of selected soil characteristics and to make decisions regarding such themes as, are the observed roots dead- or alive-, are the roots concentrated along the ped faces- or not-, what is the orientation of the stones-, etc. If figures are important, then drawing of profile sections, transects and block diagrams should be part of the basic training for a field surveyor.

Special effort should go into making the data base useful to *non-specialists to consult* the data base. Except for a chapter "Brief General Description of the Profile" no particular effort is made for this in the standard descriptions. Similar paragraphs should be provided throughout the data base, including an explanation for each soil horizon. Glossaries also help the non-specialist in collecting information from the SDB.

Acknowledgements

The author wishes to thank the Belgian National Science Foundation for a grant which allowed him to assist to the ISCOM VI Meeting.

Literature Cited

Adiwiganda, R., 1986. Classification problems of selected soil profiles from North-Sumatra (Indonesia) according to the USDA soil taxonomy system. Unpubl. M.Sc. diss., Univ. Gent.

Daniels, R.B., 1988. Pedology, a field or laboratory science? Soil Sci. Soc. Am. J. 5215, p. 1518-1519.

Estoista, R.V., 1988. Adequacy of selected profile data from Luzon (Philippines) for classification in the USDA (1975) and FAO (1974, 1987) soil taxonomic systems. Unpubl. M.Sc. diss., Univ. Gent.

FAO, 1977. Guidelines for soil profile description (2nd Ed.), 66 p, Food and Agriculture Organization of the United Nations, Rome.

FitzPatrick, E.A., 1977. Soil Description. 66p., Dept. Soil Sci., Univ. Aberdeen.

Hodgson, J.M., 1976. Soil Survey Field Handbook, describing and sampling soil profiles. Soil Surv. Techn. Monograph No. 5, U.K.

Lopulisa, C., 1986. Critical analysis of the soil characterization in soil resource inventories from the tropics, with special reference to the use of the U.S. soil taxonomy (1975) system. Unpubl. Ph.D. diss., Univ. Gent.

Lopulisa, C., Langohr, R., Msanya, B.M., 1986a. Analysis of the U.S. Soil Taxonomy (1975) for the classification of soils in the Tropics, with special reference to the Tropept suborder. Transactions XIII Congres Int. Soc. of Soil Science, vol. III, 1196-1197, Hamburg.

Lopulisa, C., Langohr, R., Msanya, B.M., 1986b. Adequacy of the existing pedon data of organic soils for the classification among the Histosols (USDA Soil taxonomy, 1975). Proc. Int. Symp. on Peat Soils (in press), Yogyakarta, Indonesia.

Soil Survey Staff, 1951. Soil Survey Manual.U.S. Dep. Agr. Handb. 18. U.S. Govt. Printing Office, Washington, D.C.

Vacca, A., 1988. Soil profile data from Italy: their adequacy for classification in the USDA (1975) and FAO (1974, 1987) taxonomies and for land suitability estimation. Unpubl. M.Sc. diss., Univ. Gent.

The Soils of the Desertic and Arid Regions of Chile

Walter Luzio-Leighton[1]

Abstract

The anticiclone from the southwest Pacific originates a high pressure system that is the main cause of the arid and desertic climate in the north of Chile. The soil moisture regime is aridic, with the exception of some close basins in the Altiplan, where it is aquic. The soil temperature regime is isothermic near the coast, thermic in the central part, and cryic in the Andes and the Altiplano.

A distinction is made between the soils from desertic conditions and those from arid and semiarid regions, and the limit in located near the parallel 29° S.L. This arbitrary boundary has been placed in the attempt to separate the desertic soils from the arid ones on the basis of the intensity and the natur of pedogenic processes that have affected both.

In the desertic region, the soils with minimum profile development are Cryorthents in the Altiplano and Torriorthents in the central "pampas" and coastal area. The soils with structure or color B horizons are Camborthids formed on finer sediments than the former. The volcanic materials show a very slight development, so that no andic properties have been identified yet. Some highly saline soils show a very hard surface crust, which makes difficult their placement in *Soil Taxonomy*.

In the arid and semiarid regions, the soils present a higher degree of development than desert soils. In the coastal area, very well developed illuviation features are found in the soils from marine terraces; thus, argillic and natric horizons are frequent. Towards the innerland there is a hilly lanscape, and the soils are not well known. Some observations indicate the presence of cambic and argillic horizons in soils derived either from limestone or from granite.

Introduction

In a first approach it is necessary to distinguish between desertic and arid or semiarid regions. In Chile, the limit is located near the parallel 29° S.L., where the annual rainfall increases from less than 100 mm to more than 200 mm but not more than 270 mm. This is an arbitrary boundary which attempts to separate desertic from srid soils on the basis of the intensity and the nature of the pedogenic processes that have affected both.

The soils of these regions are not very well known, because of the difficulties of making soil surveys and the minimal interest official and private agencies have shown, since the regions exhibits very limited farming potential. The exception is for the soils of the highly productive valleys, transverse the Intermedial Depression and approach the Pacific Ocean. Several detailed soil surveys have been made on those valleys whose principal soils are of aluvial origin; however, their significance in the total surface of arid lands is little.

[1]Walter Luzio-Leighton, Facultad de Ciencias Agrarias Y Forestales, Universidad de Chile, Casilla 1004, Santiago, Chile.

This paper provides a general outline of the soils from the desertic and arid regions in Chile, and in so doing reviews the main researches already made, and this author's unedited data. The data have been interpreted according to author's view of the distribution pattern of the soils through the most important physiographic units described in the north of Chile, from 18°S.L. to 32°S.L. (Figure 1).

Results and Discussion

Climatic Factors that Conform the Aridity in the North of Chile

The anticiclone from the south west Pacific that originates high pressure systems is located between 25° and 30°S.L. and 90° West, that is, in front of the dryest part of Chile which is normally considered from 18° to 32°S.L.

The air masses descending from the troposphere are gradually heated, even though near the ocean they are cooled, producing a thermic inversion below 1,000 m altitude. The thermic inversion restricts the vertical movements of the air masses, causing an arid climate with limited precipitation.

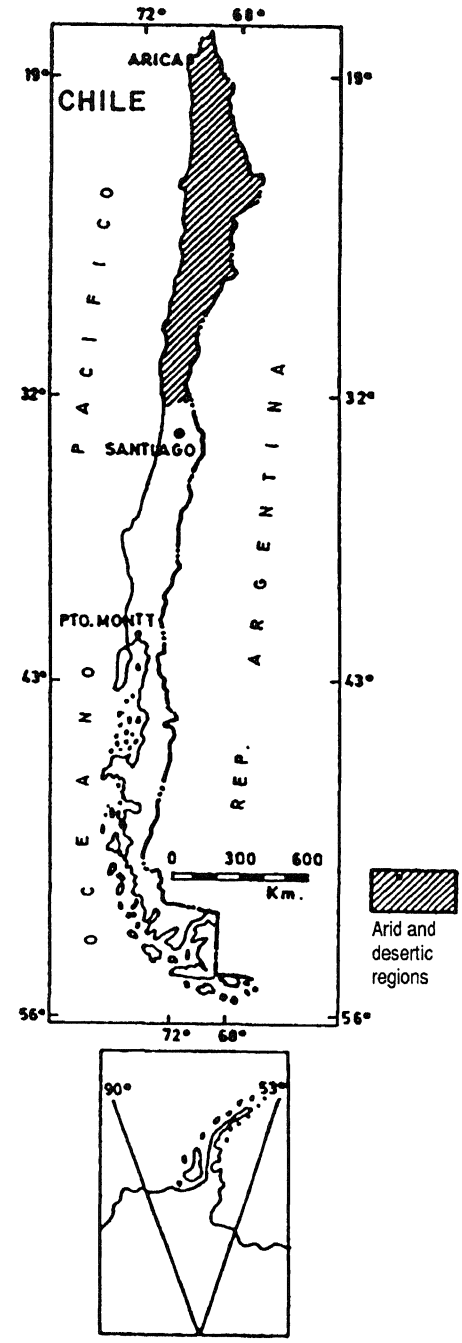


Figure 1. Arid and desertic regions of Chile and distribution area of Aridisols.

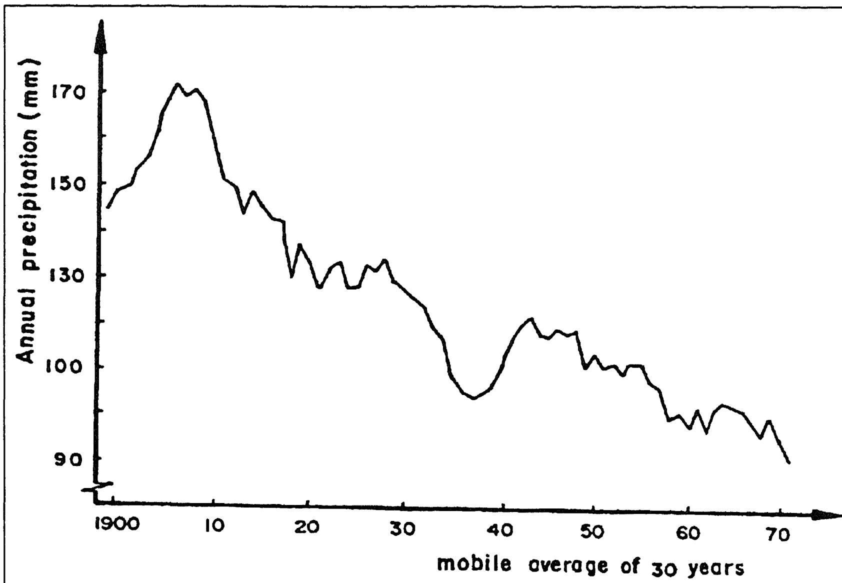


Figure 2. Rainfall tendency in La Serena (Chile) from 1900 to 1970, using the mobile average of 30 years. (Taken from Santibanez, 1985).

Considering these genetic factors of the climate, it is possible to subdivide the north region of Chile into four areas from west to east: the litoral (coastal line), the pampa (Intermedial Depression), the precordillera (Andes piedmont), and the Altiplano (Andes plateau).

Litoral Area

The litoral is a narrow strip not more than 15 km wide where the thermic conditions are alleviated by the influence of the ocean. There is no frost and the temperature varies from 7° to 28°C, considering the minimum temperature of the coldest months and the maximum temperature of the warmest months. The abundant cloudiness makes diminish solar radiation and also the thermic fluctuations. The average rainfall ranges from less 10 mm per year to 270 mm in the southern part of the litoral area. The annual moisture deficit ranges from near 2,000 mm in the northern part of the area to 1,400 mm in the southern part.

Pampa

In the second area, the pampa, the mean annual temperatures do not vary significantly from those of the litoral area, but the daily fluctuations are much greater, from 33°C during the day to lower than 0°C in the coldest nights. These conditions are dominant in the so-called "extreme desert" that is more or less up to paralel 27°S.L. Going to the south, the continental features of the climate are still present, though without desertic characteristics and approaching more mediterranean conditions. Similarly, precipitation is less than 5 mm in the desert to 250 mm in the southern part, concentrated only during winter months.

Precordillera

The third area, the precordillera extends parallel to the Andes mountains (north - south direction) as a piedmont formation at an altitude more than 2,000 but less than 2,500 m.o.s.l. The maximum temperatures are around 25°C and the minimum around 5°C. The rainfall fluctuates from 50 to 100 mm per year.

Altiplano

The fourth area, the Altiplano, is a plateau-like formation that extends over 4,000 m.o.s.l. and only up to 23°S.L. The mean temperatures are always below 6°C and frost is frequent all the year round. The climate of the Altiplano is characterized by a summer rainfall (about 200 mm) regime that produces a subhumid period lasting 1 to 3 months.

Rainfall regime tendencies

The cyclic variations of the climate at planetary scale are very well known, as is their influence on the amount and distribution of rainfalls. The desertic and arid zones of Chile are subjected to a process of climatic desertization, expressed as a clearly decreasing tendency in the amount of annual precipitation.

Figure 2 shows the rainfall tendency from 1900 to 1970, using the mobile average of 30 years technique.

Soil moisture and soil temperature regimes

Aridic regime

According to Van Wambeke and Luzio (1982), from the border with Peru up to parallel 32°S.L. there is only an aridic soil moisture regime, which goes from the sea coast to the high Andes mountains and to the Altiplano (Figure 3). Near the coast are many meteorological stations and enough data is available to apply the Newhall model of calculation, while in the Andes piedmont and the Altiplano the data are scarce and isolated. The author therefore had to assume that the aridic soil moisture regime extends to the Andes, based on data about water deficit obtained from the difference between annual precipitation and evaporation.

Aquic Regime

One of the most interesting features that characterize the Altiplano is the presence of close basins with restricted drainage where high amounts of organic matter have been accumulated, giving rise to organic soils. Similar situations are found in some depressed areas close to small streams where the water flows slowly, overflowing the surrounding terrains. There have been doubts about the use of the concept of aquic moisture regime for those cases because the water moves permanently with some dissolved oxygen, though some reduction phenomena are expressed as mottles of high chroma and gleyed colors (Luzio, 1985).

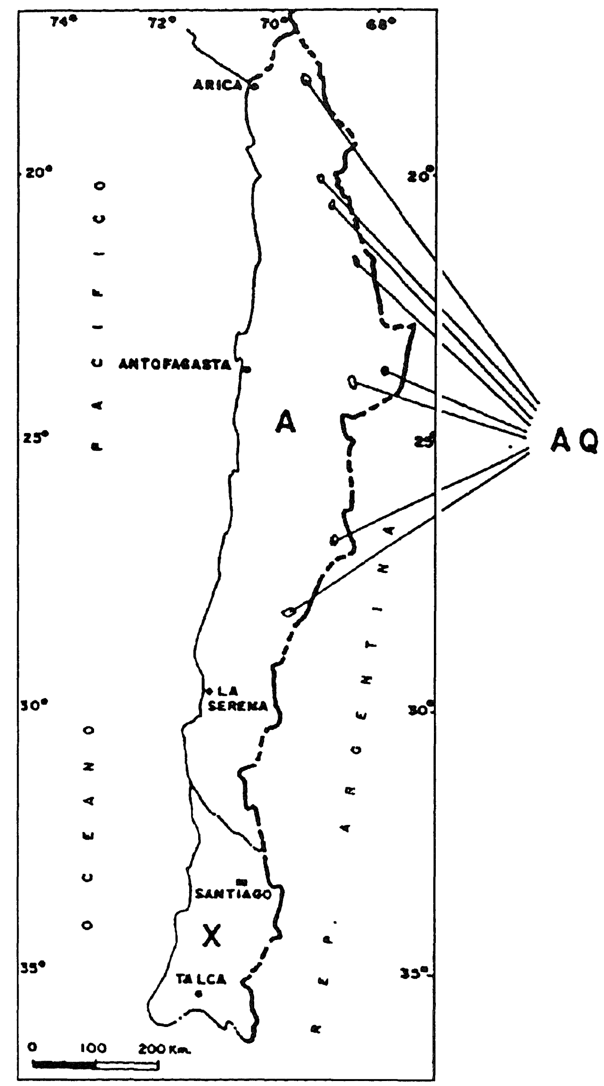


Figure 3. Soil moisture regimes in the desertic and arid regions of Chile. A: Aridic; AQ: Aquic; X: Xeric.

Iso Regimes

With regard to soil temperature regimes the iso regimes are associated with the area near the ocean. Some scarce data allow us to assume the marine influence reaches the inner area of pampas. The isothermic regime becomes the most important one from the extreme north of the country up to parallel 27°S.L. From this latitude to the south, the isothermic regime is restricted to the shore line in a strip no more than 15 km wide (Luzio, 1985).

Cryic Regime

The cryic temperature regime has been assigned to the Andes mountains and Altiplano although, as it was already pointed out, the limited or non existent data do not permit us to verify this assignment (Figure 4).

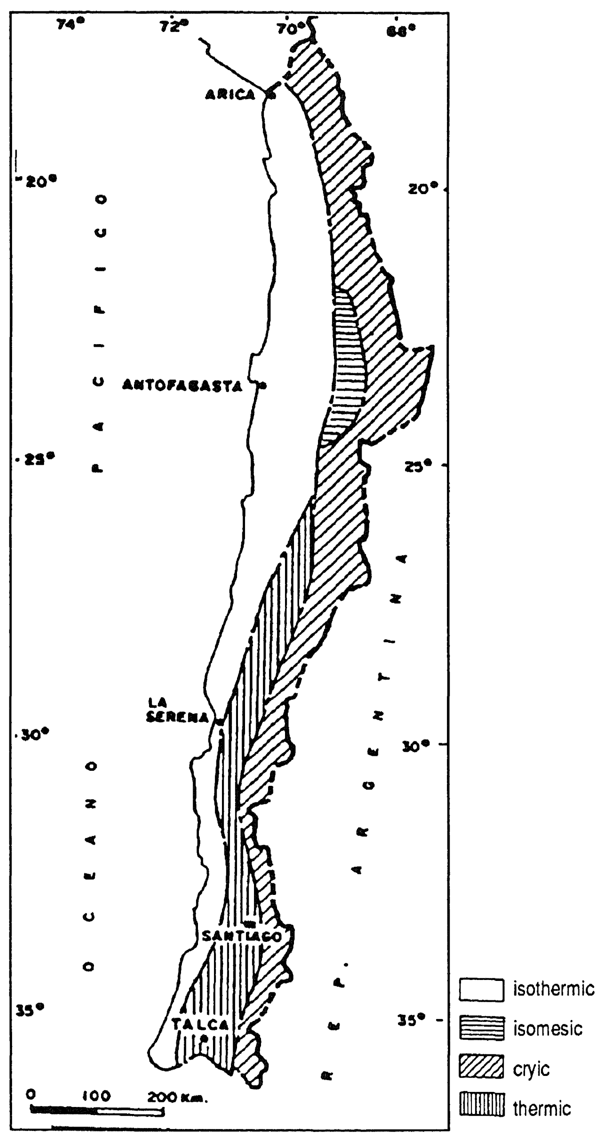


Figure 4. Soil temperature regimes in the desertic and arid regions of Chile.

The Soils

In the desert, the weathering processes have been limited to the phyiscal breakdown of the parent rocks, producing crumbled materials subjected to mass movements such as landslides or mudflows. The restricted available water produces a very weak organic regime and scarce vegetation cover and also does not allow any kind of eluviation-illuviation processes (Luzio, 1985). The conditions prevalent in the arid and semiarid regions make possible a more active organic regime, a more dense shrub cover, and the development of illuviation features enough for an argillic horizon (Luzio, 1985).

The Soils in the Desertic Region

Group 1

The first group of soils corresponds to those with a minimum profile development, such as the shallow soils over rock, stratified soils, and skeletal soils of coluvial origin. None of these soils are associated with specific landscape or area; they are found either in the coastal area, in the pampa, or in the high Andes regions and the Altiplano (Luzio and Vera, 1982). They are coarse textured soils with less than 0.2% organic carbon in the surface horizon. Those soils found in the Andes and Altiplano regions qualify as Cryorthents, and those in the pampa and coastal areas qualify as Torriorthents.

The stratified soils whose origins have to be found in aluvial transportation and sedimentation in the Andes and Altiplano show an irregular distribution of organic carbon with depth, so that we have called those soils Cryofluvents even though they are never or almost never flooded in present conditions.

Finally, in this first group, skeletal soils must be included. Normally they ahve more than 40% gravel throughout the profile, mostly sandy-skeletal or loamy-skeletal, and are associated with metastable slopes.

Group 2

A second group of soils, also found in all areas in the desert region, show some profile development in the form of a structure and/or color B horizon. The soils are formed on finer sediments than those of the first group and they have been considered as Camborthid, Lithic, and Fluventic subgroups (Alcayaga and Luzio, 1985).

Profile 1 in the Appendix is an example of this kind of soil. It is a rather homogeneous soil, without stratification and with a cambic horizon determined by the presence of soil structure and clay increase. It seems the soil is the result of its environment, where pedogenic agents have been acting according to that medium. The low soil temperatures, the limited water availability coming as infrequent rains and the almost inexistent vegetation suggest that these weakly expressed pedogenic agents have been active for a considerable period of time in order to produce a slightly to moderately developed soil. This observation agrees with Nettleton's and Peterson's ideas (1983), in the sense that the diagnostic horizons in Aridisols require considerable time to form and, consequently, Comborthids have to develop on stable surfaces of late Pleistocene or greater age.

Group 3

A third group corresponds to the soils derived from volcanic materials whose distribution is restricted only to the Altiplano area. Our experience with volcanic soils in arid and desertic regions indicates that the pedogenic development is controled completely by the arid environment and that the volcanic materials do not play a definitive role in the direction of pedogenic processes.

Some data shows that no difference can be detected in desertic soils developed from other parent materials. The organic regime is very restricted and sometimes has an irregular distribution with depth. The pH is slightly acid to neutral. A dominance of coarse sand and very coarse sand exists throughout the profile, and the majority of roots are concentrated in the second horizon. The percentage of glasses measured in all the sand fractions is between 40 to 60%, the P retention is lower than 15 percent and Al + 1/2Fe is generally lower than 0,15 percent and only in some subhorizons is slightly higher than this value. These figures indicate that the desertic conditions are responsible for the very low degree of weathering of the volcanic glasses. Depending on the presence or absence of a structure B horizon, they have been considered as Camborthids or Cryorthents. No one of the few profiles analized up to now meet the requirements of the andic properties according to Circular Letter N° 10 (INCOMAND, 1988).

Moreover, considering our present knowledge about these kind of soils, we do not think a Vitric subgroup in Orthids would be useful, because the behavior of the soils will not differ significantly from the Typic ones (Profile 2 is an example of this kind of soil).

Group 4

The fourth group are the soils with high organic matter content developed only in closed basins and depressed areas of small streams in the Altiplano. It seems the moisture regime is a reducing one, at least in some periods of the year, even though the water table is normally moving water. There is no frost and the mean annual soil temperature is higher than 0°C.

These areas, called "Bofedales," are not great in extent but they are very important as grassland resource that maintains the livestock of the region. The classification of these soils has been subjected to some discussion because of the moving water table. Some of them qualify as organic soils (Fibrist and Hemist) and others, with smaller organic matter content, as Cryaquents and Cryaquepts, based on the presence of gley horizons and oxidation-reduction mottles throughout the profiles.

The pedon from Turi (Profile 3) corresponds to a stratified soil with high calcium carbonate content and alkaline reaction throughout the profile. The soil has been formed from medium particle size materials in a somewhat poorly drained closed basin. During rainy periods the ground water table may reach the soil surface and the rest of the year is not deeper than 80 cm.

Group 5

The last group of soils in the desert region corresponds to those highly saline soils found only in the Intermedial Depresion, at altitudes near 1,000 m.o.s.l.

The most striking feature of these soils is the presence of a very hard surface crust of salts about 25 cm thick. Few analyses from the crust and from the different strata have been made, and they show the dominant cation is Na+ and the anion is Cl-; Ca++ and SO_4— are of secondary importance. The surface crust is composed also by 50 to 65 percent clay (Luzio and Vera, 1982). Our knowledge is very limited about this kind of soil, only a small number of profile descriptions and analyses have been made, indicating the topsoil is an indurated saline horizon composed mainly of NaCl.

These kinds of horizons have not been described yet in *Soil Taxonomy*. More studies are needed, so that at present we only can ask if we should have to consider a "petrosalic" horizon.

The Soils in the Arid Regions

The arid regions extend from 27°S.L. approximately to 32°S.L. The soils show a higher degree of development in comparison to the desert soils. The presence of cambic B horizon is common, and argillic horizons are not rare on the marine terraces.

Soils from the Coastal Range

The region is characterized by the presence of numerous marine terraces of different levels. In those places where there are no marine terraces, the Coastal Batholith reaches the shore line and the soils are formed "in situ" from the weathering of the plutonic rocks (Luzio et.a., 1978).

Significant areas of the terraces are formed by stabilized sand dunes covered with xerophytic shrubs and scarces Acacia trees. The lack of development of these soils is attributed to the very coarse and recent sediments, unstable sur-

faces due to mass movements or erosion, and the lack of enough water to promote the soil processes. They have been considered mostly as Torripsamments (Alcayaga and Luzio, 1985).

In higher level terraces from the Pliocene (Munoz, 1950; Paskoff, 1970), soils with argillic horizons are normally developed. The presence of illuviated horizons has been interpreted as formed on stable surfaces in a more humid climate than present conditions with pronounced rainy and dry seasons. Paleargid is the most frequent great group.

The existence of wetter climates during the Pliocene and the Quaternary has been emphasized by different authors who give diverse kinds of arguments, such as the presence of large aluvial terraces and aluvial cones, the presence of relic forest, and so on.

In some medium terraces, argillic horizons are replaced by natric horizons whose presence is attributed to the numerous marine transgressions due to epeirogenic movements. The accumulation of Na+ has taken place along with Mg++ so that some of the sampled natric horizons show 32 percent Na saturation and nearly 60 percent Mg saturation. They are mostly Xerollic Natragids (Alcayaga and Luzio, 1985).

Finally, in the marine terraces, soils over calcareous sediments are found. The more important are those with a petrocalcic horizon formed from a high amount of marine bivalved molluscans shells clearly recognizable in the lower part of the profiles.

Innerland soils

The knowledge of the soils of this area is very preliminary because no systematic pedogenic studies or soil inventories on any scale exist.

Limestone is one of the significant parent rocks found not only in the flat relief of the intermontane valleys but also in the hilly topography. At present there is no a clear idea about the origin of the lime in the hilly topography, because most of the hills are built up by plutonic rocks. Argillic horizons are developed over a calcic horizon and no petrocalcic horizons have been found yet.

Other important parent rock is granite. The soils developed from this rock are supposed to be Camborthids.

Conclusions

1. The soils formed from volcanic materials are found clearly in the Andes, under a cryic temperature regime, and from 18° to 27° SL, approximately. They are Cryorthents. It is thought that the majority of the other soils have some kind of contamination with volcanic materials, though this is very difficult to evaluate.
2. In the desertic region there is a clear dominance of young soils (Torriorthents) without significant pedogenic processes or diagnostic horizons other than an ochric epipedon.
3. In the desertic region the most intense soil development is a cambic horizon. The soils in the Intermedial Depression are Typic and Lithic Camborthids and those in the Andes, with cryic temperature regime, are also Camborthids. No subgroups are now recognized.
4. Paleorthids are found both in the desertic and the arid region.
5. The soils with high organic matter content (Histosols or not) are found only in the high mountains in close basins with impeded drainage. The most feasible explanation for their formation is associated with the postglacial pond cycle.
6. Argillic horizons are found only in the arid regions and associated with the soils developed on marine terraces. They are Paleargids.
7. Natric horizons are only found in restricted areas in the arid region on marine terraces. The high exchageable Na+ and the high Mg++ have been interpreted as a decisive influence of pleistocene marine transgressions.
8. The valleys are dominated by non-developed soils from aluvial origin: Torriorthents, Torrifluvents, and Torripsamments. Some valleys have a high salinity status.

Literature Cited

Alcayaga, S., and W. Luzio. 1985. Clasificaci ón taxonómica de los suelos de regiones des érticas y àridas del norte de Chile. Bolet ìn N° 5 (2): 141 - 144. Sociedad Chilena de la Ciencia del Suelo, Santiago, Chile.

INCOMAND, 1988. Circular Letter N 10. 29 February, 1988. 80 p.

Luzio, W. 1985. Gènesis y clasificaci ón de los suelos de regiones àridas y deserticas de Chile. Boletin N ° 5(1): 107 - 140. Sociedad Chilena de la Ciencia del Suelo, Santiago, Chile.

Luzio, W., I. Badilla, and W. Vera. 1978. Zonificacion del sistema fitoed àfico en el secano costero de la IV Region. II Simposio Nacional de la Ciencia del Suelo. Santiago, Facultad de Agronom ìa, Universidad de Chile: 476-495.

Luzio, W., and W. Vera. 1982. Caracterizaciòn de suelos en ecosistemas de la I y la II Region. SACOR - CODECIAGRO. Informe para la CORFO. 126 p.

Munzo, C. Jorge. Geolog ìa. En: Geografìa Econòmica de Chile. Tomo I.Corporacion de Fomento de la Producciòn. 55 - 87.

Paskoff, 1970. Le Chili semi-aride. Recherches geomorphologiques. Viscaye Freres, Paris. 420 p..

Santibanez, F. 1985. Rasgos agroclimáticos de la zona árida chilena. Boletin de la Sociedad Chilena de la Ciencia del Suelo N ° 5: 1 - 28. Santiago, Chile.

Van Wambeke, A., and W. Luzio. 1982. Determinacòn de los reg ìmenes de humedad y temperatura para los suelos de Chile. Agricultura T écnica (Chile) 42(2): 149 - 159.

Appendix

Profile 1.

Location: 30.8 km from San Pedro de Atacama, in the road from San Pedro de Atacama to Tatio.

Altitude: 3.750 m.o.s.l.

Geomorphic position: footslope

Slope: 5 - 10%

Stoniness: 60% angular gravel, 5% boulders

Classification: Typic Camborthid

A 0 - 11 cm. Dark reddish brown (5 YR 3/3) and brown (7.5 YR 5/4, dry); loamy sand; single grain; loose; non plastic; non sticky; few fine roots; 10% gravel; clear smooth boundary.

A2 11 - 23 cm. Dark reddish brown (5 YR 3/3) and brown 7.5 YR 5/4, dry); sandy loam; weak fine subangular blocky structure; slightly hard; friable; slightly plastic, non sticky; common fine roots; few fine pores; 5% gravel; clear smooth boundary.

B1 23 - 45 cm. Dark reddish brown (5 YR 3/3) and brown (7.5 YR 5/4, dry); sandy clay loam; weak fine and medium subangular blocky structure; slightly hard, friable; plastic, slightly sticky; common fine and medium roots, 3% gravel; clear smooth boundary.

B2 45 - 55 cm. Dark reddish brown (5 YR 3/3) and brown (7.5 YR 5/4, dry); sandy clay loam; weak fine subangular blocky structure; hard; friable; plastic, slightly sticky; common fine roots; few fine pores; 30% gravel; abrupt wavy boundary.

R 55 - 75 cm Tuff.

Profile 1: Typic Camborthid (Tolar S.P. Atacama - Tatio pedon).

Depth. (cm)	Hor.	pH	O.C. %	Particle size distribution (%) sand	silt	clay
0 - 11	A1	5.0	0.34	72.0	16.0	12.0
11 - 23	A2	7.0	0.02	66.0	18.0	16.0
23 - 45	B1	7.0	0.33	56.0	18.0	26.0
45 - 65	B2	6.9	0.27	60.0	20.0	20.0
Depth. (cm)	**Hor.**	**Extractable cations Ca**	**Mg**	**Na**	**CEC K**	**B.S. _%**
		meq/100 g				
0 - 11	A1	2.4	0.6	0.3	0.7	_6.3 63
11 - 23	A2	4.8	1.9	0.4	0.9	_8.9 89
23 - 45	B1	7.2	3.0	0.6	1.6	12.5 99
45 - 65	B2	7.2	3.2	0.4	1.7	13.0 96
Depth. (cm)	**Hor.**	**Water content (%) 1/3 BAR**	**15**	**mmho/cm**	**E.C.**	
0 - 11	A1	12.7	6.8	0.6		
11 - 23	A2	13.6	7.9	0.2		
23 - 45	B1	17.4	11.0	0.5		
45 - 65	B2	18.0	11.5	0.7		

Profile 2.

Location: 15 km from Putre town, in the road from Putre to Chungara.

Altitude: 4,300 m.o.s.l.

Geomorphic position: Altiplanic plateau

Slope: 5 - 7%

Coarse fragments: 5%

Erosion: Probably hydric and eolic

Classification: Typic Cryorthent.

A1 0 - 8 cm. Very dark grayish brown (10 YR 3/2) and dark reddish brown (5 YR 3/2.5, dry); coarse sandy loam; weak fine and medium subangular blocky structure; friable; plastic, non sticky; few fine roots; common fine pores; 30% gravel; clear smooth boundary.

A2 8 - 15 cm. Variegated, reddish (5 YR 4/4) as dominant; silty loam; with gravels; very fine granular structure; friable; slightly plastic, sticky; many fine roots; many fine pores; 45% gravel (1 cm); clear smooth boundary.

A/C 15 - 21 cm. Variegated, yellowish red (5 YR 4/6) as dominant; silty clay loam with gravels; weak fine granular structure; slightly plastic, slightly sticky; few fine roots; 60% gravel (up to 4 cm); clear smooth boundary.

C 21 - 115 cm. Variegated, light brown and brown (10 YR 7/3 and 10 YR 5/3) as dominant; gravelly sandy loam; massive; very hard, indurated, no roots, no pores; 70% pumice gravel (2 - 4 cm).

Observations: rhyolitic pumice fragments are abundant throughout this profile. They correspond partially to the gravel described.

Profile 2: Typic Cryorthent (Putre pedon).

Depth. (cm)	Hor.	pH	O.C. %	Particle size distribution (%) sand	silt	clay
0 - 8	A11	7.0	0.29	56.0	24.0	20.0
8 - 15	A2	7.2	0.63	52.0	20.0	28.0
15 - 21	A/C	7.3	0.46	47.2	24.8	28.0
21 -115	C	7.1	0.1	60.0	22.0	18.0

Depth. (cm)	Hor.	Extractable cations Ca	Mg	Na	K	CEC	B.S. %
		meq/100g					
0 - 8	A11	3.6	1.9	0.3	0.5	8.8	72
8 - 15	A2	7.5	3.3	0.4	1.0	15.6	78
15 - 21	A/C	10.4	4.4	0.5	1.2	20.0	83
21 -115	C	8.6	3.6	0.6	1.0	13.7	100

Depth. (cm)	Hor.	Water content (%) 1/3 BAR	15	E.C. mmho/cm
0 - 8	A11	9.5	5.7	0.3
8 - 15	A2	16.5	10.8	0.2
15 - 21	A/C	21.6	15.3	0.3
21 -115	C	20.5	12.6	0.3

Profile 3. Typic Cryaquent, calcareous (Turi pedon).

Depth. (cm)	Hor.	pH	O.C. %	$CaCO_3$ %	Particle size distribution (%) sand	silt	clay
0 - 15	A	9.0	6.0	21.8	31.4	48.0	20.6
15 - 22	C1	9.2	8.1	55.6	32.6	48.8	18.6
22 - 29	2C2	9.1	4.1	70.4	67.4	22.0	10.6
29 - 66	3Cg	9.1	6.5	__.__	43.4	34.0	22.6

Depth. (cm)	Hor.	Extractable cations Ca	Mg	Na	K	CEC
		meq/100 g				
0 - 15	A	79.2	14.2	121.8	3.89	30.0
15 - 22	C1	33.1	13.0	___3.79	2.81	32.2
22 - 29	2C2	28.3	8.2	___9.0	1.5	22.8
29 - 66	3Cg	27.1	4.0	___6.5	1.5	14.1

Depth. (cm)	Hor.	Water content (%) 1/3 BAR	15	ESP _%	E.C. mmho/cm	Soluble cations Ca	Mg	Na	K
0 - 15	A	56.5	25.2	6.6	150	6.6	15.7	22.2	7.6
15 - 22	C16	78.5	36.1	10.1	__14	1.4	_1.4	12.0	1.8
22 - 29	2C2	57.9	25.9	__.__	___6	_.__	__.__	__.__	_.__
29 - 66	3Cg	52.8	28.7	__.__	___4	_.__	__.__	__.__	_.

Profile 3.

Location: "Vega de Turi", between Toconce and Aiquina villages

Altitude: 2.900 m.o.s.l.

Geomorphic position: extended closed basin

Slope: 2%

Water table: 66 cm depth

Calcareous surface crust

Classification: Typic Cryaquent, calcareous

A 0 - 15 cm. Brown to dark brown (7.5 YR 4/4); silty clay loam; weak medium subangular blocky structure; friable; plastic, slightly sticky; many fine and medium roots; many fine pores; oxidations about 30% of the horizon; violent effervescent (HCl); clear smooth boundary.

Cl 15 - 22 cm. Reddish yellow (7.5 YR 7/8); silty loam; massive; friable; plastic and sticky; common fine and medium roots; many fine pores; violent effervescent (HCl); abrupt smooth boundary.

2C2 22 - 29 cm. White (10 YR 8/2) with charcoal layers, 5 mm thick, of black color (2.5 Y 2/0); silty clay loam; massive; firm; plastic and sticky; few fine and medium roots; common fine pores; violent effervescent (HC1); abrupt smooth boundary.

3Cg 29 - 66 cm. Gray (5 Y 5/1) intermixed with black (2.5 Y 2/0) layers of charcoal; gravelly clayey loam; massive; plastic and sticky; few fine and medium roots; many fine pores; violent effervescent (HCl).

Observations: ground water level: 66 cm

A Review of Recent Research on Swelling Clay Soils in Canada

A.R. Mermut,[1] D.F. Acton,[1] and C. Tarnocai[2]

Abstract

This paper reviews recent studies of the characteristics, behavior, genesis, and classification of swell-shrink soils in Saskatchewan. All the soils studied crack to varying degrees and have distinct slickensides in the subsoil. The solum of the soils is much deeper than originally thought. Horizon designation is a problem in these soils. A case for B horizon can be argued, but this B horizon should be recognized as Bw. The identification of C horizons may be based on structure and the depth of gypsum and salts.

Physical, chemical, and mineralogical properties of the soils vary along an environmental gradient from south to north. Micromorphological features reflect the magnitude of the stress and soil displacement. Stress features were minimal or nonexistent in the C horizons. Genesis of the soils is governed by the smectite content (extent of swelling), climate, and vegetation. While the majority of swell-shrink soils may be considered as Vertisols, these soils would not fit into any of the Vertisol suborders of *Soil Taxonomy*. Therefore, a new suborder, Borerts, is appropriate for the Vertisols occurring in frigid and cryic soil temperature regimes. Currently, swell-shrink soils are classified at the family level in the Canadian system. It seems inconsistent with the differentiating criteria used at higher categories in the Canadian system to relegate these criteria to a lower categorical level. The Canadian system could be improved by developing a separate order for swelling soils.

Introduction

Heavy clay soils with swell-shrink properties occur in glacial lake sediments that extend from sub-arid grassland to subhumid grassland-forest transitional zone in the Canadian Prairies. Early soil survey work in Saskatchewan (Mitchell et al., 1944) considered clay soils as a separate type from other regional soils because of their high fertility status and their resistance to drought. At the national level, these soils were variously referred to as Argillaceous Regosols or Grumic soils, and they have been classified at various categorical levels. Clay soils with marked swell-shrink potential were not separated from others at the order level in the Canadian System of Soil Classification (CSSC). However, grumic subgroups were created within the Chernozemic order (Clayton, 1963) to accommodate these soils. These subgroups remained in the system until 1974 (Canada Department of Agriculture, 1974). However, in the final version of the CSSC (Canada Soil Survey Committee, 1978), "grumic" property has been used to separate the soils at the family level. In the CSSC the term "grumic" is intended to indicate those soils that have fine texture, smectitic mineralogy, and self-mulching properties. It can be argued that self-mulching properties are related to soil genesis; therefore, these soils should be considered above the family level. The lack of data on swell-shrink properties of clay soils likely hindered the use of these criteria in soil classification in Canada.

While it has been long recognized that clay soils differ from other Chernozemic soils in the region, no attempt was made to study them in sufficient detail to understand their genesis, properties, and relations to Vertisols described in *Soil Taxonomy* (Soil Survey Staff, 1975). Soils that have morphological characteristics similar to Vertisols but occur in colder regions were excluded from this order. Thus, until very recently, the soils in the north central U.S.A. and the Canadian Prairies, with frigid and cryic soil temperature regimes, were not considered within the Vertisol order. In September 1980 approval was given by USDA Soil Conservation Service to drop the temperature requirement for this soil order (Soil Survey Staff, 1982).

Several research studies were initiated to examine the characteristics of these soils in Saskatchewan. The objective of the present paper is to review the recent information regarding the morphology, mineralogy, micromorphology, and soil forming processes that are operative in swell-shrink soils in Saskatchewan and to evaluate these in terms of their genesis and classification at the international levels.

[1]Saskatchewan Institute of Pedology, University of Saskatchewan, Saskatoon, SK S7N 0W0 Canada. Contribution No. R627.

[2]Agriculture Canada Land Resource Research Centre, Ottawa, Ontario K1A 0C6, Canada.

Morphology and Micromorphology

Horizon Designation

The lack of color contrast, common occurrence of carbonates throughout the profile, and poor expression of structure when moist, make horizon designation difficult in the heavy clay soils in Saskatchewan (Mermut and St. Arnaud, 1983; Dasog et al. 1987). However, coarse prismatic, breaking to angular blocky structure, may be observed when the soils are dry. In Canada, if the horizon below the Ap is calcareous or massive, such a horizon is designated as a C horizon (Ayres et al., 1985). Therefore, designating the solum and differentiating this from the C horizons is often difficult (Mermut and St. Arnaud, 1983). The boundary between the solum, which may be more than one meter thick, and the C horizon is often sharp, although it may be diffuse and a transitional horizon may be recognized. The upper part of the C horizon is physically disrupted and, at depth, prominent varying with partial disruption may be observed. However, at the contact zone between the solum and C horizon, more soluble salts and associated gypsum often are concentrated, which may be helpful in determining the depth of solum. As will be discussed later, micromorphological studies are helpful in determining the horizon designation as well as in understanding the genesis of these soils.

Another question is how to designate the horizon within the solum. Subsoil horizons that show compound prismatic-blocky structure when dry are considered to be a B horizon (Canada Soil Survey Committee, 1978; Soil Management Support Services, 1982). Dasog et al. (1987) suggested that a case for B horizon can be argued on structural consideration alone and that these B horizons should be recognized as Bw (Bmk in Canadian System). Wilding and Tessier (1988) indicated that, based on morphological and other characteristics, it is fully appropriate to recognize the presence of a "cambic horizon" in Vertisols.

Cracking and Slickensides

All heavy clay soils in Saskatchewan show cracking. Despite its importance for classifying Vertisols, information on direct measurement of crack parameters is rare in the literature. Limited studies by Dasog (1986) showed that cracking in the subarid regions of Saskatchewan is less than one half as intense as in Vertisols in the xeric moisture regime of Israel (Yaalon and Kalmar, 1978) and even lower than in the Vertisols of Sudan (El Abedine and Robinson, 1971). In both Israel and Sudan, summers are much drier than in Saskatchewan and precipitation during the summers in Saskatchewan effectively decreases the intensity of cracks.

Under native grassland, cracks are shallower and narrower. The moisture regime of Saskatchewan soils influences the duration and intensity of cracking and limited information suggests that cracks may remain open for more than 90 cumulative days. Considering the moisture distribution gradient, one would expect a decreasing intensity from south-west to north-east. Furthermore, days are much longer during the summer and, therefore, more effective in crack development. This fact needs to be considered while comparing these soils with other swell-shrink soils that occur in southern latitudes. The degree and frequency of changes in moisture content of the soil are perhaps the most important parameter that controls the intensity of cracking and movement within the soil, provided that soils have sufficient COLE (enough swelling clays) to produce movement. The degree of swelling and shrinking is decreased with organic matter, carbonates, gypsum, high electrolyte concentration, sesquioxides, and low activity clays which bind and cement the soil fabric.Dasog et al. (1988) found the following relationship between COLE and the content of expandable clay in Saskatchewan soils (1):

COLE = -0.0026 + 0.0033 x % expandable clays (1).

This would mean that, to produce COLE value of 0.1 > 30%, swelling clay is needed. This equation agrees well with that reported by Schafer and Singer (1976) for some swell-shrink soils in California (2):

COLE = -0.00123 + 0.00336 x % expandable clays (2).

Another equation suggested by Dasog (1986) indicates the relationships between fine clay and COLE (3):

COLE = 0.042 + 0.0037 x % fine clay (3).

According to Grossman et al. (1985), a soil with an intermediate COLE value of 0.05, at 1,500 kPa moisture content, would produce cracks of about 1 cm wide and repeat distance between the cracks of about 20 cm. If we apply an intermediate COLE value of 0.05 to the above equations, a minimum of 15% swelling clay or 17% fine clay would be required. These will approximately equal a minimum of 30%

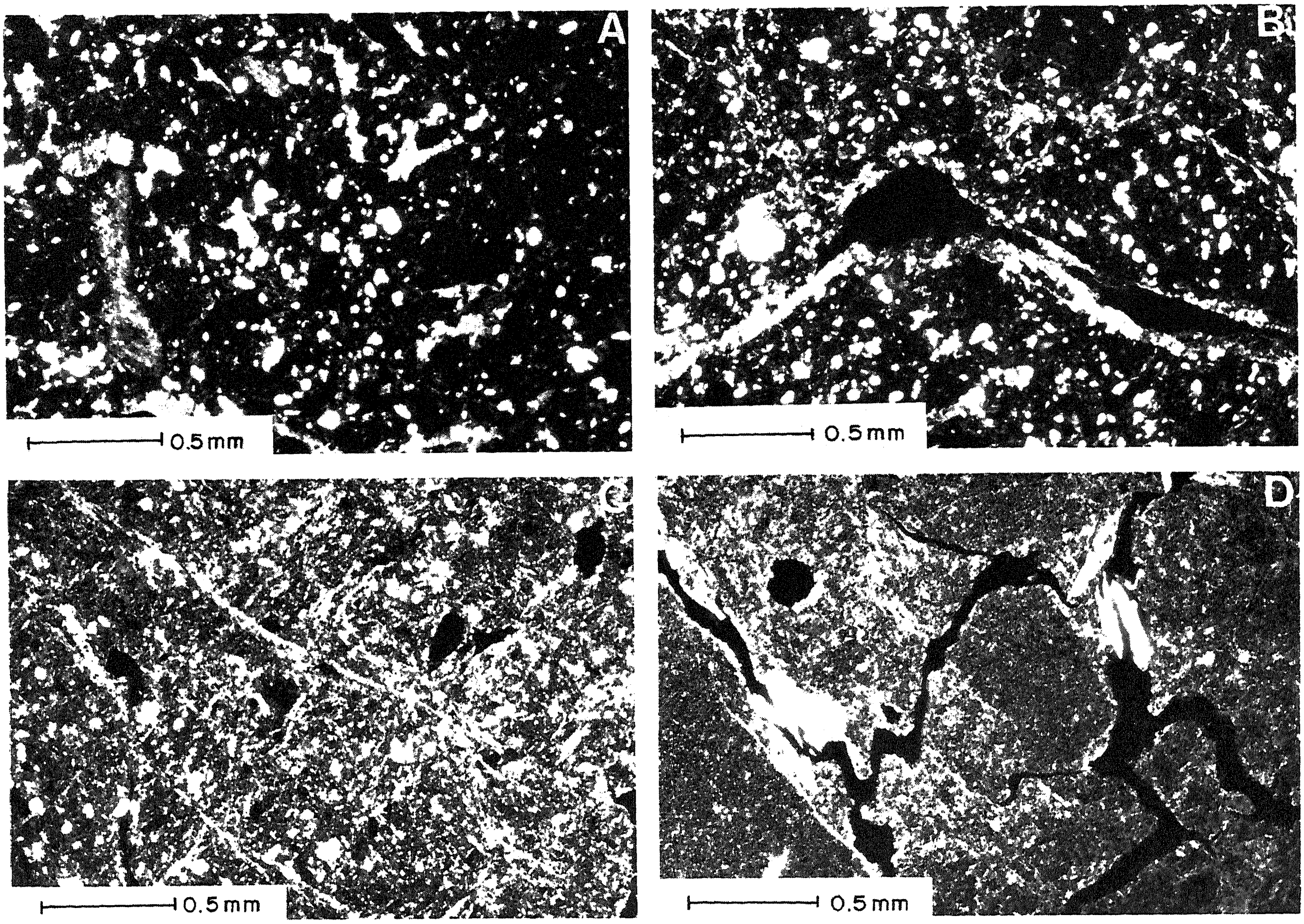


Fig. 1 Photomicrographs from the Melfort soil. A. mullgranoidic (A1p1 horizon 2-10 cm). B. ferriargillans (Bnt horizon 23-31 cm). C. parallel striated b-fabric in very dense mass (Bw1 horizon, 42-50 cm). D broken peds with striated fabric and planar voids.

total clay, which is also equal to the clay content criterion used by Soil Survey Staff (1975) to differentiate Vertisols. We therefore suggest that there is no need to mention clay content in the definition, but a definition of diagnostic vertic features is now desirable.

Slickensides occur in the subsoil of all swell-shrink soils studied in the four major soil zones of Saskatchewan, where the difference between horizontal and vertical stresses is large. It appears that, because of the lower overburden pressure, the upper part of the solum remains relatively stable. The intact columnar and prismatic structures of the shallow Solonetzic soil formed on the heavy Melfort and Tisdale soils and the presence of slickensides in the heavy subsoil below the Solonetzic B horizon testify to this view.

Micromorphology

Swelling heavy clay soils in Saskatchewan have distinct micromorphological features (Mermut and St. Arnaud, 1983; Dasog et al. 1987). Because of cracking and the formation of slickensides, these soils have produced entirely different void patterns (Figs. 1, 2). Subparallel joint planes, slickensides, meta-skew planes, and craze planes with smoothed surface conformation are the types encountered (Mermut and St. Arnaud, 1983). Both cracking and soil displacement produces planar voids. Micromorphology provides an excellent opportunity to differentiate the voids that are produced by stress from those formed by simple desiccation. At depth, surfaces of planes are generally slickensides (Fig. 2C), indicating the high stress and displacement resulting from the swelling clays.

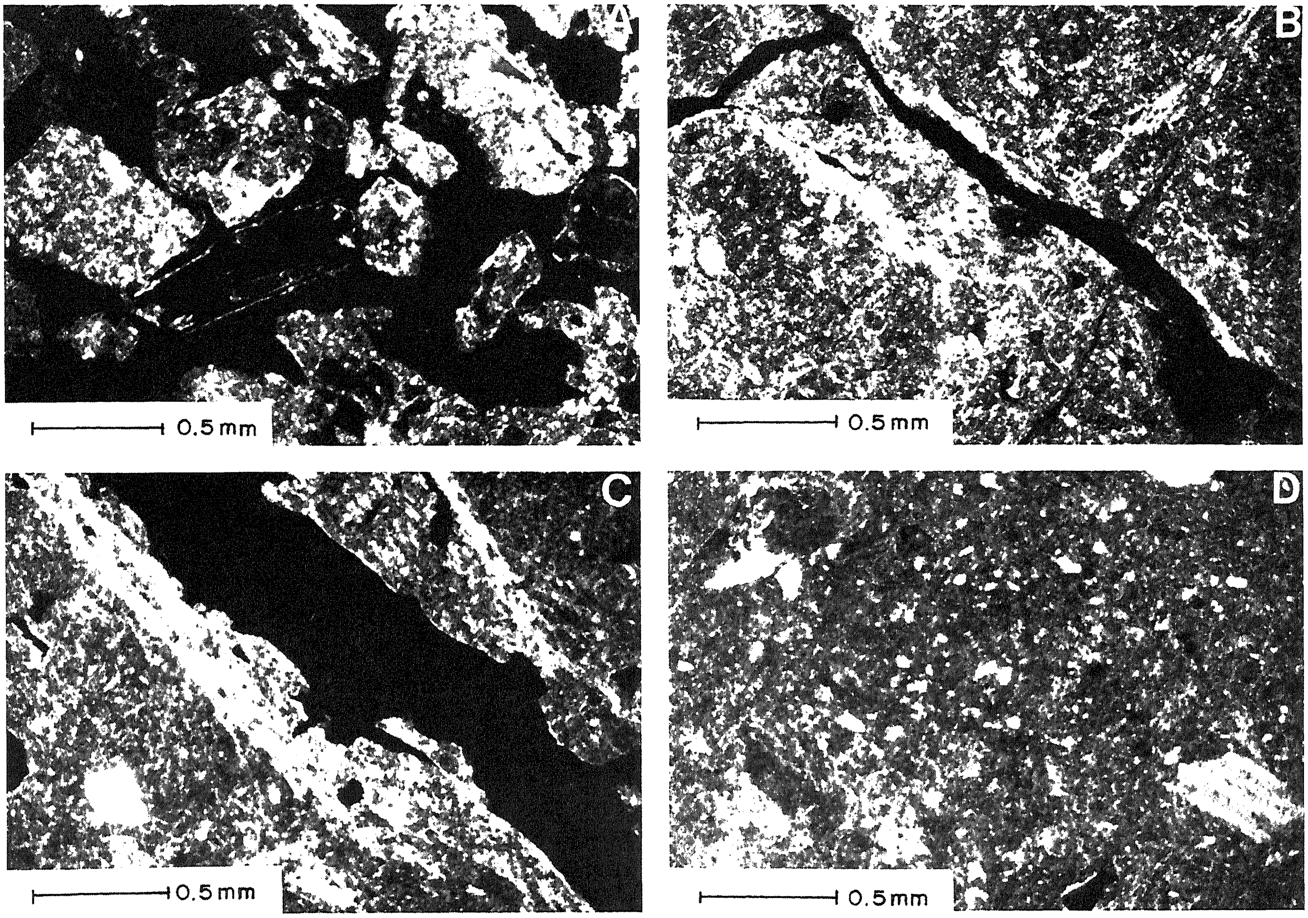


Fig. 2 Photomicrographs from the Kelvington soil. A. surface granular structure (Ap horizon, 3-13 cm). B reticulate striated b-fabric (AB horizon 18-28 cm) and planar voids. C. porostriated b-fabric (Bw horizon, 38-48 cm). D. physically broken parent material without any orientation (argillasepic, porphyroskelic) (BC horizon, 110-120 cm).

Because of shrinkage tension in the soil material, a decrease of plasma volume occurs and this produces the system of cracks and very dense ground mass (Figs. 1C, 2B) (Porphyroskelic related distribution pattern in Brewer, 1976 terminology).

The clay soils have stress, related masepic fabric (striated b-fabric according to Bullock et al., 1985) below the surface soil, and this typical plasma separation (Figs. 1C, 2B) extends down to about 1 m, where the original glacio-lacustrine sediments are found to be highly broken, maintaining their original unistrial fabric (sedimentary fabric) (Fig. 1D). The proportion of unistrial fabric as measured by Dasog et al. (1987) increases with depth (Fig. 3). In the Sceptre, Regina, Melfort, and Tisdale soils, all horizons designated as C contained more than 60% unistrial fabric, suggesting that moisture change frequency is minimal at this depth. Moisture measurement by de Jong and McDonald (1975) showed that little change in moisture occurred at a depth of 105-135 cm on a Sceptre clay. The transitional BC horizons have 30-60% unistrial fabric that can be differentiated from the rest of the solum. This transitional zone also contains gypsum, which can be used as an additional characteristic in identification of this zone.

Although micromorphology helps in precisely differentiating the transitional and C horizons, it may not be the most convenient and practical technique for horizon designation. A large seasonal moisture deficit may be responsible for better expression of masepic fabric as well as for the formation of slickensides. The higher amounts of iron-manganese nodules observed also suggest that masepic fabric is better ex-

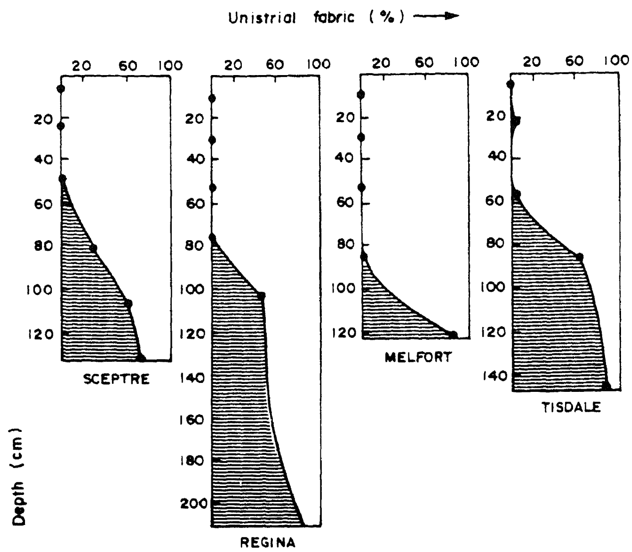


Fig. 3 Occurrence of unistrial fabric in the soils studied, estimated by point counts (from Dasog et al. 1987).

pressed in the Rouleau and Kelvington soils. These two soils remain wet for a certain period of time. Desiccation of the soil, during an unusually dry summer, likely creates very high tension, causing the formation of large cracks.

Ap horizons display dominantly matrigranic-metamatrigranoidic fabric (associated with granular structure, Fig. 2A). As this type of fabric also is found in other soils, such as Chernozemic and Luvisolic soils, this characteristic alone cannot be used to differentiate Vertisols from other soils.

Soil nodules, with sharp and diffuse boundaries, embedded in the soil matrix are other striking features found in the clay soils. The number of nodules decreases with increasing depth. As will be discussed later, they possess strong evidence of the churning process.

Mineralogy

Several investigators (Warder and Dion, 1952; Rice et al., 1959; Kodama and Brydon, 1965; Mills and Zwarich, 1972; Dudas and Pawluk, 1982) have reported the occurrence of smectite and illite as the predominant clay minerals in the soils of the prairie provinces of western Canada. Detailed characterization of some swelling lacustrine parent material from southern Saskatchewan (Mermut et al., 1984) confirmed that they are dominated by smectite and have a similar mineral composition. This suggests that the clays originate from the same source. The general order of clay mineral abundance, which remains almost constant throughout the profile, is smectite (> 50%), illite (15-30%), kaolinite (10%), vermiculite (< 10%), and quartz (< 10%). Smectites in Saskatchewan have high iron and their chemical composition are similar to smectites of Indian Vertisols (Mermut et al., 1984).

Dasog (1986) reported that vermiculite contents may be significant in swelling soils occurring in the northern subhumid grassland-forest transition and forest regions and, therefore, these soils are expected to have a higher layer charge. The degree of stress-induced plasma separation in the soils of the transitional zone is much higher than other swelling soils in Saskatchewan. Several factors play a role in the development of soil plasma. Because of the low swelling abilities of vermiculites (due to their high charge), they form very distinctly oriented clay domains under stress. Southern Saskatchewan soils have more organic matter in the surface horizons, which may have a masking effect on plasma separation. Contrarily, due to low organic matter in the subsoil, swell-shrink potential is likely higher in the northern soils. Furthermore, large fluctuations in the soil moisture state of northern soils may be responsible for a high magnitude of volume change that ultimately produces a high degree of plasma separation.

The soils have up to 50% fine clay and have a high surface area (600-800 $m2g^{-1}$), important factors when considering the swell-shrink potential of these soils. cole values range from 0.1 to 0.168, except for the surface horizons of the Melfort and Tisdale soils that have high contents of organic matter. Southern clays contain more than 15% exchangeable Na+ and high amounts of electrolytes at depth and the amounts of exchangeable Na and electrolytes decrease gradually towards the north. It is clear from the foregoing that, at this stage, it is very difficult to predict which soils are more prone to soil displacement.

As indicated by Wilding and Tessier (1988), in the presence of high electrolyte, only interparticle pore water contributes to field volume change. If the electrolyte concentration is low and smectites are saturated with Na, both interparticle and a portion of the interlayer water contribute to field volume changes with changing soil moisture content. High electrolyte concentration in this regard may be a factor that reduces the swelling pressure in the BC and C horizon. Furthermore, this zone also contains

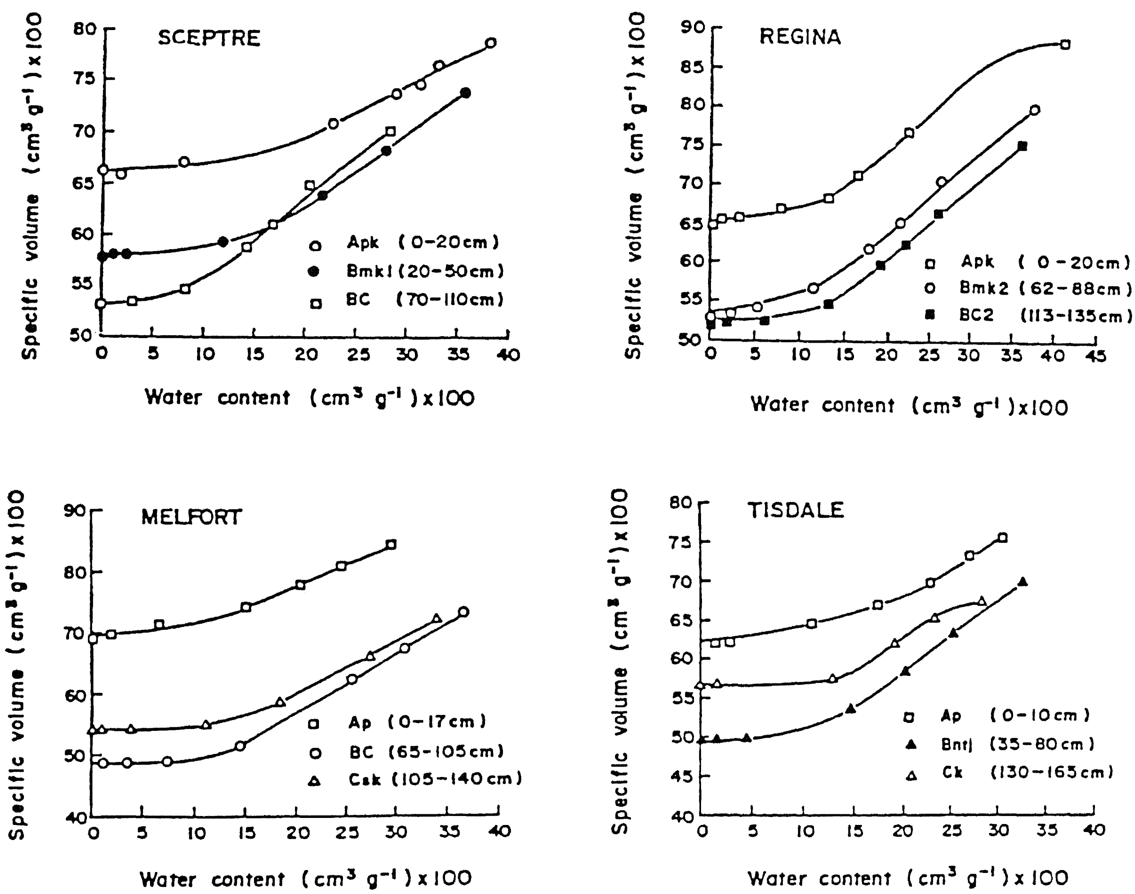


Fig. 4 Shrinkage curves of selected horizons from four clay soils in Saskatchewan (from Dasog, 1986).

appreciable amounts of gypsum. It is known that gypsum independently promotes flocculation of clays and lowers swell-shrink potential.

Shrinkage curves of some selected horizons of the Sceptre, Regina, Melfort, and Tisdale (Fig. 4) show that there is a shrinkage in the range of field moisture;therefore cracking is expected in these soils.

Soil Genesis and Classification

Soil Genesis

Swelling clays are found in three climatic regions in Saskatchewan: sub-arid, semi-arid, and sub-humid. However, the lack of information does not allow us to discuss the genesis of the soils in different climatic regions. The discussion provided here is more general and reflects our current knowledge.

As indicated above, all clay soils in Saskatchewan crack as a result of seasonal moisture fluctuation. It is postulated that the degree of the moisture deficit is higher for the sub-arid and semi-arid Sceptre and Regina than for the sub-humid Melfort, Tisdale, and Kelvington soils. As indicated by Soil Survey Staff (1975), the soils that are in vertic subgroups have potential for movement but may not become dry enough or moist enough to produce sufficient movement.

In a laboratory study, Fredlund (1975) observed swelling pressures of about 400 to 1000 kPa and shear strengths of 20 to 40 kPa when Regina soil was moistened. This shows that the smectite containing clays have high swelling pressure and low shear strength and, therefore, the potential for soil displacement and formation of slickensides exists in these soils. As indicated above, slickensides are observed in the

subsoil in all four major soil zones in Saskatchewan. This, according to Mermut and Acton (1985), shows that heavy clay soils undergo sufficient cyclic cracking and swelling in most years to produce appreciable soil movement; hence the slickensides.

Several mechanisms were suggested regarding the formation of slickensides (Ahmad, 1985; Wilding and Tessier, 1988). It is not the intent of this paper to elaborate on these mechanisms, but rather to suggest some ideas about the possible pathways that result in the formation of slickensides in Saskatchewan soils. More than one cycle of wetting and drying is possible each year, once during the spring melt and once or twice during the summer. It is known that cracks promote bypass flow of water, which induces differential wetting in the subsoil. Dasog et al. (1987) suggest that such wetting occurs during the periods of heavy rain in the late summer or fall. In a year when fall precipitation is low, the soil may remain cracked during winter. In such years, differential wetting occurs due to bypass flow of meltwater. During this flow of water, in addition to crack infill by gravity, some surface fine granular structural aggregates also are transported. Tonguing of the surface soils and the presence of soil nodules, decreasing with increasing depth, are strong evidence of downward movement of surficial material. Such a movement also was considered as evidence of the churning process. While recent reports (Wilding and Tessier, 1988) suggest that the crack infilling process is only partially functional in the genesis of Vertisols, it is a process that makes the Vertisols distinctly different from other soils.

Our limited observation revealed that the slickensides in the subsoil are less frequent and they all have a similar nature. They are deepest and most frequent in the Rouleau soil. This is expected, as this soil is much deeper than other swell-shrink soils and it has a soil moisture regime that can be classified as aquic. Despite the presence of well preserved columnar and prismatic structure and clay cutans in the B horizon of the Melfort and Tisdale soils (Fig. 1A, B), the occurrence of the slickensides in the subsoil raises several serious questions. It appears that slickensides form in the subsoils of these two soils by vertical overburden and horizontal swelling pressures. Displacement forces and associated features generated in the subsoil do not cause any major disturbance in the upper part of the solum, and this part remains relatively stable. Yaalon and Kalmar (1978) reported that low overburden pressure and cracks would prevent high horizontal stress. Dudal and Eswaran (1988) identified several distinct zones or horizons in Vertisols. They recognize that surface soils, subject to cracking, have large prisms which may part to coarse, angular blocky elements. Slickensides were never found within the upper 20 cm of the Vertisols.

While the above discussion may confirm that pedoturbation is an unimportant process in the Melfort and Tisdale soils, it suggests that the influence of subsoil displacement on the upper solum may be lesser than in the soils of sub-arid and semi-arid regions (Sceptre and Regina). Relatively thicker soils observed in depressional areas in the Regina clay plain may explain the surficial rearrangement of material in these swell-shrink soils (Mermut and Acton, 1985). In addition, these studies on Saskatchewan soils show that the features related to swelling and shrinking are not as distinct as in comparable subtropical and tropical soils.

Soil Classification According to *Soil Taxonomy*

The swell-shrink soils in Saskatchewan, described above, meet the requirement of the new concept of Vertisol as described in the 5th circular letter by Comerma (1989) and can, therefore, be classified within the Vertisol soil order. Earlier, Vertisols that have frigid or colder soil temperature regimes were excluded in SMSS *Soil Taxonomy* for unknown reasons (Comerma et al., 1988). The soils do not meet the criteria of any suborders of the Vertisol described by Soil Survey Staff (1975). The only suborder that is closest to this group of soils is Xererts. However, clay soils in Saskatchewan do not have xeric moisture regime for two reasons: 1) cracks may open and close more than once each year, and 2) the xeric moisture regime is that typified in Mediterranean climates. Therefore, based on soil temperature regime, a new suborder, "Borerts," equivalent to Borolls and Boralfs, is proposed (Dasog et al., 1987). By definition, Borerts are Vertisols that have frigid or cryic soil temperature regimes. This is slightly different than that proposed by Comerma et al. (1988) in which they exclude frigid temperature regime from the definition.

In the 5th circular letter by Comerma (1989) it is recommended (J. Witty) that the name Borert be changed to Cryert. Temperatures reported by Treidl (1979) suggest that the tem-

perature regime of the southern Saskatchewan soils is frigid and that of northern soils is cryic (Boreal and Cryoboreal as defined in Canada). By definition, the soil temperature regime of Boreal environment is frigid or cryic, but some have a pergelic temperature regime (Soil Survey Staff, 1975). This means that a range of cold temperature regimes is included in Boreal. Therefore, it seems more appropriate to keep the term Borert so that the soils that have both frigid and cryic soil temperature regimes can be recognized under one suborder.

There are an estimated 14560 km2 of the Sceptre, Regina, and similar soils in the Brown and Dark Brown soil zones alone in Canada (Clayton et al., 1977) and appreciable areas within the Black (Mermut and St. Arnaud, 1983) and Gray soil zones of Saskatchewan and Manitoba that may meet the definition of Borerts. The swelling clays in the Red River Valley of North Dakota, the Grumusols of Montana (Hogan et al., 1967), and gilgaied soils in South Dakota (White and Bonestell, 1960) also may be included in this suborder.

Perhaps two great groups, Haploborerts and Humiborerts, can be recognized in Borerts. Soils such as Kelvington have a high content of organic matter, and should be recognized and classified as Humiborerts. We have made this proposal to the International Committee and we hope that the meeting will enable us to clarify this point. By definition, Humiborerts have 12 kg or more organic C in a unit volume of 1 square meter to a depth of 1 meter below the top of the mineral soil surface, exclusive of any O horizon that may be present.

Several subgroups have been suggested in Comerma's 5th circular letter (1989). Our observations have shown that soils in sub-arid and semiarid regions of Saskatchewan have one or more horizons with an ESP of $\geq$15 or a SAR of $\geq$13 within a depth of 100 cm of the soil surface and meet the requirement of "sodic" subgroup (Table 1). According to Wilding and Tessier (1988), clays may be dispersed at relatively low SAR and ESP (5 to 8% or less) under low electrolyte systems and an ESP of 15% or more may be required in high electrolyte system. Poorly drained soils occurring in depressional areas, that are saturated with water for some time, can be classified as "aquic." Soils in the Kelvington area have a moist color value of 4 or a dry value of 6 or more in one or more subhorizons within the upper 30 cm of the soil surface, reflecting the influence of parent material.

These soils have low organic matter and very high clay contents. The dark soils that have a moist color value of 3 or less, and chroma of 2 or less, and a dry color value of 5 or less in all subhorizons within the upper 30 cm of the surface, previously grouped in the "pell" categories, will be called "typic." Considering the 5th circular letter by Comerma (1989), the following three subgroups are suggested within Haplo and Humiborerts:

Sodic Haploborerts are Borerts that have within a depth of 100 cm of the soil surface one or more subhorizons with an ESP of $\geq$ 15 or SAR of $\geq$ 13.

Chromic Haploborerts are Borerts that have a dominant value, moist, of 4 or more and a chroma of 3 or more or a value, dry of 6 or more in some part of the upper 30 cm of the soil surface.

Typic Haploborerts, other Haploborerts.

Definition of Vertic Horizon

As indicated above, a definition of a diagnostic vertic subsurface horizon, such as Bw, is needed. Slickenside formation, as a result of soil displacement, is an essential process for Vertisols. The definition, therefore, should be centered around the features related to soil displacement. Soils that show slickensides and wedge shaped aggregates generally are deeper and have enough swelling clay to promote cracking during the dry season. It seems that slickenside and wedge shaped aggregates are the most essential features for Vertisols. Comerma et al. (1988) suggest that gilgai should no longer be a required criterion at the order level to classify a soil as a Vertisol.

Using the existing criteria for Vertisol (Soil Survey Staff, 1982), the following simple definition can be made. A vertic horizon is a subsurface horizon or horizons with slickensides close enough to intersect, or wedge shaped peds that

Table 1. SAR and exchangeable Na% in the Rouleau and Kelvington soils.

Rouleau Depth cm	Exchangeable SAR	Na%	Organic carbon%	Kelvington Depth cm	SAR	Exchangeable Na%	Organic carbon %
0-15	1	1	1.59	0-10	-	1	3.00
15-29	1	1	1.18	10-31	RR	1	0.80
29-46	4	3	1.07	31-65	1	2	0.48
46-71	11	9	0.84	65-99	1	2	0.40
71-97	15	10	0.51	99-131	2	3	0.35
97-133	21	8	0.42	131-183	2	3	0.33
133-163	14	7	0.49	183-210	1	3	0.23
163-200	1	1	0.57				

have their long axes 10 to 60° from the horizontal. A vertic horizon is 25 cm or more thick, with its upper boundary within 100 cm of the soil surface.

Proposed Revisions in the Canadian System

Although the number of representative soils observed and analyzed in Saskatchewan may not be warrant to propose changes in the Canadian classification system. As it stands, the following proposal is an outline to form the basis for establishing the need for changes in the system. More work is needed to determine the extent and distribution of the swelling clay soils in Alberta and Manitoba.

The soils that have diagnostic vertic subsurface horizons may be classified within a new "Vertisolic" order in the Canadian System of Soil Classification. There are presently two fundamental problems within the classification of clay soils: 1) there are examples of these soils occurr in sub- and semi-arid regions which do not meet color criteria of a Chernozemic A horizon and 2) considering the evidence for a B horizon, because of the presence of carbonates, these soils also should not be classified in the Regosolic order. It is also difficult to reconcile "grumic" as a family criterion among other recognized criteria, as this criterion is related to genesis but not properties and has to be recognized at a higher level. This suggestion for a revision is consistent with the philosophy of the system, whereby classes at higher categorical levels reflect the broad differences in the soil environments that are related to the differences in soil genesis (Canada Soil Survey Committee, 1978).

Using the same principles of the CSSC System, four great groups can be established:

Brown, Dark Brown, Black, and Dark Gray (including Gray).

Within each group the following subgroups can be recognized:

Orthic, Rego, Calcareous, Solonetzic, Gleyed, and combination of these.

Most soils in Saskatchewan that have high shrink-swell potential have cracks for some period in most years and a potential LE of 6 cm in the upper 1 m. These soils that have potential but do not exhibit the required characteristics should be recognized within the Vertic subgroup, such as Vertic Dark Brown Chermozem.

References

Ahmad, N. 1985. Acid Vertisols of Trinidad. p. 141-151. In Proceedings of 5th Int. Soil Correlation Workshop, Soil Survey Administration, Ministry of Agriculture and Irrigation, Khartoum, Sudan.

Ayres, K.W., D.F. Acton and J.G. Ellis. 1985. The soils of the Swift Current map area, 72J Saskatchewan. Extension Publ. No. 481 University of Saskatchewan, Saskatoon.

Brewer, R. 1976. Fabric and mineral analysis of soils. Robert E. Kriger Publishing Co., Huntington, N.Y.

Bullock, P., N. Fedoroff, A. Jongerius, G. Stoops and T. Tursina. 1985. Handbook for soil thin section description, Waine Research Publ., Wolverhampton, England.

Canada Department of Agriculture. 1974. The system of soil classification for Canada, Publ. 1455, Ottawa, Canada.

Canada Soil Survey Committee. 1978. The Canadian system of soil classification. Can. Dep. Agric. Pub. 1646, Supply and Services, Canada, Ottawa.

Clayton, J.S. 1963. Report on classification of Chernozemic soils. In report on the fifth meeting of the National Soil Survey Committee of Canada, 4 - 8 March. Winnipeg, Manitoba. National Soil Survey Committee, Agriculture Canada, Ottawa.

Clayton, J.S., W.A. Ehrlich, D.B. Cann, J.H. Day, and I.B. Marshall. 1977. Soils of Canada, Vol. II Soil inventory. Research Branch, Can. Dept. Agric. Ministry of Supply and Services Canada, Ottawa.

Comerma, J.A. 1989. Fifth circular letter International Committee on Vertisols. USDA-SCS, Washington, D.C.

Comerma, J.A., D. Williams and A. Newman. 1988. Conceptual changes in the classification of Vertisols. p. 41-54. In L.P. Wilding and R. Puentes (ed.) Vertisols: Their distribution, properties, classification and management. Texas A and M University Printing Center, College Station, TX 77843.

Grossman, R.B., W.D. Nettleton, B.R. Brasher. 1985. Application of pedology to plant response prediction for tropical Vertisol. p. 97-116. In Soil Survey Administration Khartoum Sudan (ed.), Proceedings of the 5th International Soil Classification Workshop - Taxonomy and Management of Vertisols and Aridisols, Part I papers 2-11 November, Khartoum, Khartoum University Press, Khartoum, Sudan.

Dasog, G.S. 1986. Properties, genesis and classification of clay soils in Saskatchewan. Ph.D. Thesis, University of Saskatchewan, Saskatoon, Canada.

Dasog, G.S., D.F. Acton and A.R. Mermut. 1987. Genesis and classification of clay soils with vertic properties in Saskatchewan. Soil Sci. Soc. Am. J. 51: 1243-1250.

de Jong, E. and K.B. McDonald. 1975. The soil moisture regime under native grassland. Geoderma 14: 207-221.

Dudal, R. and H. Eswaran. 1988. Distribution, properties and classification of Vertisols. p. 1-22. In L.P. Wilding and R. Puentes (ed.) Vertisols: their distribution, properties, classification and management. Texas A and M University Printing Center, College Station, TX 77843.

Dudas, M.J. and S. Pawluk. 1982. Reevaluation of the occurrence of interstratified clays and other phyllosilicates in southern Alberta soils. Can. J. Soil Sci. 62: 61-69.

El Abedine, Z. and G.H. Robinson. 1971. A study of cracking in some Vertisols of Sudan. Geoderma 5: 229-241.

Fredlund, D.J. 1975. Engineering properties of expansive clays. Internal. Research Report (IR-7), Transportation and Geotechnical Group, Department of Civil Engineering, University of Saskatchewan.

Hogan, E.K., J.L. Parker, V.K. Haderlie, R.C. McConnell, and W.W. Janssen. 1967. Soil Survey of Judith Basin area, Montana-series 1959, no. 42. USDA-SCS. U.S. Government Printing Office, Washington, D.C.

Kodama, H. and J. E. Brydon. 1965. Interstratified montmorillonite-mica clays from subsoil of the Prairie Provinces, Western Canada. Clays Clay Miner. 13: 151-173.

Mermut, A.R. and D.F. Acton. 1985. Surficial rearrangement and cracking in swelling clay soils of the glacial lake Regina basin in Saskatchewan. Can. J. Soil Sci. 65: 317-327.

Mermut, A.R. and R.J. St. Arnaud. 1983. Micromorphology of some Chernozemic soils with grumic properties in Saskatchewan, Canada. Soil Sci. Soc. Am. J. 47:536-541.

Mermut, A.R., K. Ghebre-Egziabhier, and R.J. St. Arnaud. 1984. The nature of smectites in some fine textured lacustrine parent materials in southern Saskatchewan. Can. J. Soil Sci. 64: 481-494.

Mills, J.G. and M.A. Zwarich. 1972. Recognition of interstratified clays. Clays Clay Miner. 20: 169-174.

Mitchell, J., H.C. Moss and J.S. Clayton. 1944. Soil survey of southern Saskatchewan. Soil Survey Rep. No. 12, University of Saskatchewan, Saskatoon, Canada.

Rice, H.M., S.A. Forman and L.M. Party. 1959. A study of some profiles from major soil zones in Saskatchewan and Alberta. Can. J. Soil Sci. 39: 165-177.

Schafer, W.M. and M.J. Singer. 1976. Influence of physical and mineralogical properties on swelling of soils in Yolo county, California. Soil Sci. Soc. Am. J. 40: 557-562.

Soil Management Support Services. 1986. Designation for master horizons and layers in soils. Department of Agronomy, College of Agriculture and Life Sciences, Cornell University.

Soil Survey Staff. 1975. Soil Taxonomy: A basic system of soil classification for making and interpreting soil surveys. USDA-SCS. Agric. Handb. 436. U.S. Government Printing Office, Washington, D.C.

Soil Survey Staff. 1982. Amendment to Soil Taxonomy, part 615. In National Soil Taxonomy Handbook issue No. 1. 430-VI-NSTH, 1982, USDA-SCS, Washington, D.C.

Treidl, R.A. 1979. Handbook on agricultural and forest meteorology, Part II. Fisheries and Environment Canada, Ministry of Supply and Services, Ottawa, Canada.

Warder, F.G. and H.G. Dion. 1952. The nature of the clay minerals in some Saskatchewan soils. Sci. Agric. 32: 535-547.

White, E.M. and R.G. Bonestell. 1960. Some gilgaied soils in South Dakota. Soil Sci. Soc. Am. Proc. 24: 305-309.

Wilding, L.P. and D. Tessier. 1988. Genesis of Vertisols: Shrink-swell phenomena. p. 47-62. In L.P. Wilding and R. Puentes (ed.) Vertisols: Their distribution, properties, classification and management. Texas A and M University Printing Center, College Station, TX 77843.

Yaalon, D.H. and D. Kalmar. 1978. Dynamics of cracking and swelling clay soils: displacement of skeletal grains, optimum depth of slickensides, and rate of intrapedonic turbation. Earth Surface Processes, 3: 31-42.

Thermal Regime and Morphology of Clay Soils in Manitoba, Canada

G.F. Mills,[1] R.G. Eilers,[2] and H. Veldhuis[2]

Abstract

Soil thermal regime currently takes precedence over physical features such as continuous cracking and slickensides in the classification of clayey soils in the Vertisol order in *Soil Taxonomy.* Thermal regime data from a transect of nine clayey soils from latitude 49°N to 56°17'N were used to evaluate the extension of Cryerts (cold Vertisols) to northern latitudes. Soil and vegetation conditions range from Black and Dark Gray Chernozems (Typic Cryoboroll and Argic Cryoboroll) under subhumid grassland and grassland-forest transition to Gray Luvisols (Cryoboralf) in the cool, more humid, Boreal forest. Soil thermal regimes vary from cold to very cold. The most northerly Luvisols have permafrost within 1.5 m, resulting in a very cold thermal regime. Mean annual and mean summer soil temperatures at 50 cm decrease northward at a rate of 0.9°C and 1.5°C per degree of latitude, respectively.

These soils have the clay content and mineralogy with potential to develop vertic soil properties. Although vertic features were not noted on the transect, observations of closely comparable soils throughout the same regions indicate a decreasing presence of vertic features from southern Black soils to the northern Luvisols. Vertic properties are sufficiently common in the Black and Dark Gray soils for their classification as Cryerts if the temperature limit is altered to include colder soils. Cold to very cold Luvisols and very cold Cryosols in combination with a generally higher micaceous clay content, more humid soil moisture regime, and cryoturbation effects have less potential for development of vertic soil features. Consequently, vertic properties are not sufficiently common or well developed in Luvisol or Cryosol soils to warrant classification in the proposed Cryert suborder.

Introduction

Manitoba is centrally located in the mid latitudes (49oN to 60oN) of North America. The ecological zonation across this latitude ranges from grassland and grassland-forest transition vegetation with Chernozemic Black (Cryoboroll) soils in the south, to Boreal forest associated with Luvisolic (Cryoboralf) soils in central regions, and open Subarctic forest and Tundra associated with Cryosolic (Pergelic Ruptic Cryochrept) soils in the north. Clay soils cover some 7,000 kha (Canada-Manitoba Soil Survey, 1989) between latitude 49°N and 58°N (Figure 1).

The objective of this paper is twofold: 1) to describe the thermal regime of well to moderately well drained clay soils under forest cover as determined from nine benchmark sites along a south-north transect, and 2) to evaluate the possible extension of Cryerts (cold Vertisols) in northern latitudes.

Geological Setting

Physiographically and geologically the study area transects two large, distinctly different areas: old, massive, Precambrian crystalline rock forming the Canadian Shield in the north and younger, mainly sedimentary rock of the Manitoba Plain and the Saskatchewan Plain in the south (Bostock, 1970). The entire area was subjected to multiple glaciation during the Pleistocene Ice Age (Prest, 1970).

During the final recession of the continental ice sheet, extensive areas were inundated by glacial Lake Agassiz ponded between higher lands to the south and the receding ice front. Extensive areas of clayey-textured sediments in the Shield region occur as level lacustrine basins and as undulating and hummocky plains where the bedrock is closer to the surface. The soils range in drainage from well to poor in undulating to hummocky terrain and poorly to very poorly drained in flat lying terrain. In the Manitoba Plain, lacustrine clay deposits blanket the level to gently undulating morainal deposits in which low-lying areas are dominantly imperfectly and poorly drained.

The lacustrine clays of both regions are fine clayey and dominantly moderately calcareous. Sediments in the Manitoba Plain are derived primarily from Cretaceous shale bedrock which underlies the Saskatchewan Plain to the west.

[1]Soils and Crops Branch, Manitoba Agriculture.

[2]Land Resource Research Center, Agriculture Canada, 362 Ellis Building, University of Manitoba, Winnipeg, MB, R3T2N2.

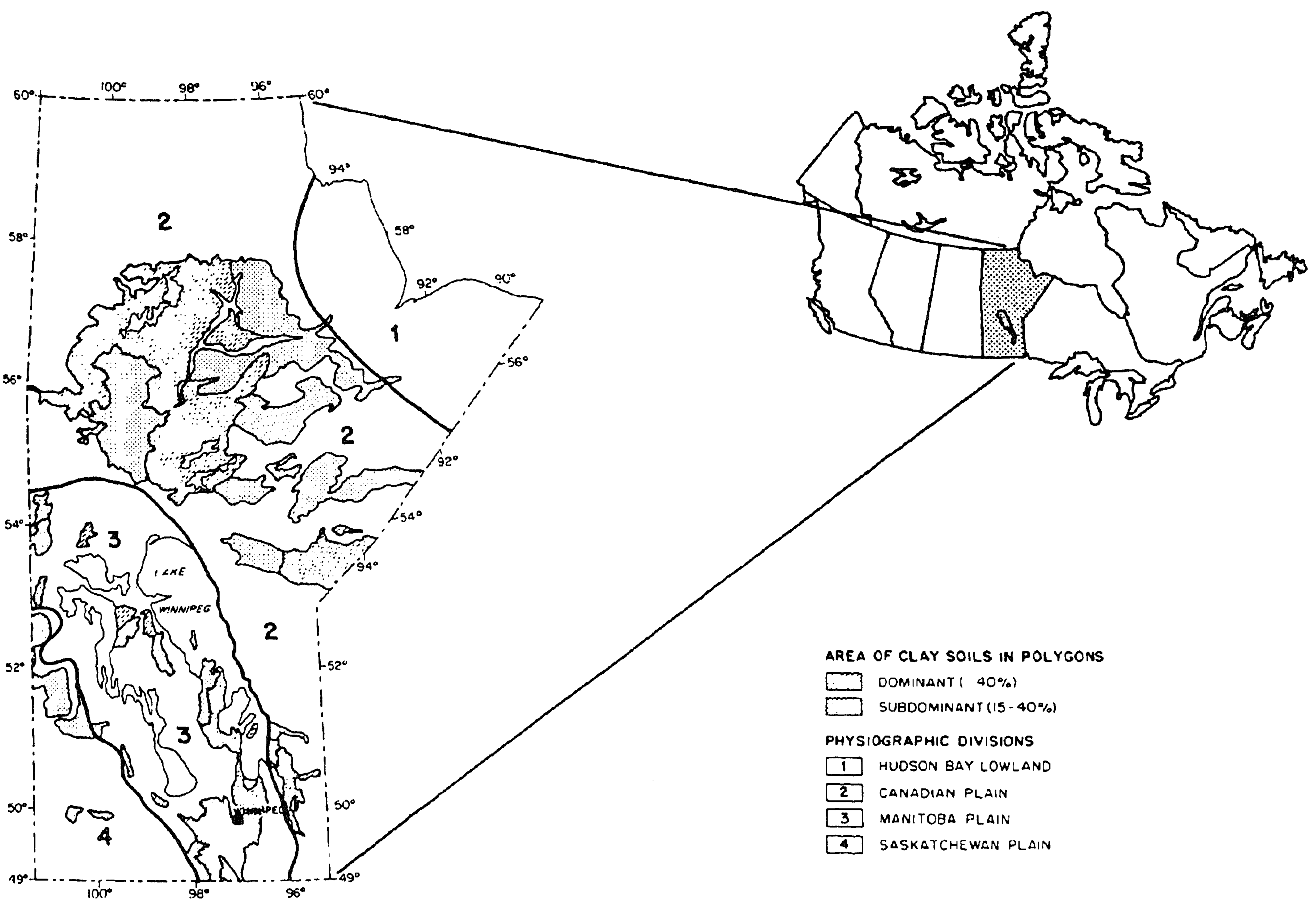


Figure 1. Physiographic divisions and extent of clayey soils in Manitoba.

In contrast, sediments in the Shield consist of clays derived from both Cretaceous rock and the local Shield rock. Although clay mineralogy is mixed in both regions, high shrink-swell clays such as smectite are co-dominant with mica (illite) in the south, and illitic clay derived from the Shield rock is dominant to the north of Lake Winnipeg. (J.G. Madden, personal communication).

Climatic Conditions

Manitoba, due to its location in interior North America at mid latitudes, is characterized by a continental climate with relatively short, warm summers and long, cold winters, with continuous snow cover from about November 20 to March 30 in the south and from about October 30 to May 1 in the north. The climatic data along a south-north transect through the study area show a gradual cooling and a decrease in length of frost-free season with increasing latitude (Table 1). At the north end of Lake Winnipeg, climatic conditions result in the first occurrence of permafrost in organic deposits. Permafrost is characterized by some depth in the soil where the temperature will remain below 0°C over at least two consecutive winters and the intervening summer (Brown and Kupsch, 1974).

Although total precipitation generally decreases to the north, the atmospheric moisture balance along the transect shifts from seasonal summer deficit in the south to very short periods of deficit or slight surplus in the north. The change in moisture deficit results from reduced evapotranspiration due to the generally cooler and shorter growing season in the north.

Ecoclimatic Regions

Ecoclimatic regions are relatively homogenous areas of the Earth's surface characterized by distinctive ecological responses to macroclimate as expressed by vegetation, soils, fauna, and aquatic systems (Ecoregion Working Group, 1989). Broad climate-soil-vegetation relationships are described in terms of seven major

Table 1. Climatic data from selected Manitoba weather stations (Atmospheric Environment Service, 1980).

Station	Latitude north	Longitude west	Mean air temperature (°C) Annual	Jan	July	Mean Frost-free period (days)	Precipitation (mm) Annual	May-Sept.	Period of continuous snow cover*
	Grassland transition Region, Gt								
Emerso	49°00'	97°16'	3.1	-18.2	17.5	125	515.7	343.6	
Winnipeg	49°54'	97°14'	2.1	-19.3	19.6	122	525.5	350.2	Nov. 21 - March 30
Stonewall	50°07'	97°20'	1.1	-20.3	19.0	118	538.23	38.9	
	Low Boreal subhumid Region, LBs								
Gypsumville	51°40'	98°44'	0.7	-20.7	18.0	101	418.4	271.2	Nov. 23 - April 12
	Mid Boreal subhumid Region, MBs								
Grand Rapids	53°11'	99°16'	-0.7	-21.8	17.7	101	442.7	267.9	Nov. 15 - April 22
	High Boreal subhumid Region, HBs								
Wabowden	54°55'	98°38'	-2.2	-24.6	16.9	105	464.2	297.6	Oct. 31 - April 28
Thompson	55°48'	97°52'	-3.9	-26.6	15.6	63	542.4	330 3	Oct. 29 - April 21
	Low Subarctic Region, LS								
Gillam	56°21'	94°42'	-4.6	-26.9	15.1	60	422.3	278.8	Oct. 30 - May 1

* Average date of persistent snow cover based on first date for 2 cm of snow cover lasting 7 days and last date of 2 cm of snow cover for last continuous 7 days. Winnipeg Climate Centre, Unpublished data from National Climatological Archive.

ecoclimatic regions. Figure 2 shows the locations of these regions and Table 2 provides a brief summary of their characteristics. (Manitoba Ecoclimatic Region Working Group, 1985).

Materials and Methods

Location and Description of Study Transect

Figure 2 shows the locations of the nine benchmark sites on the south-north transect. Table 1 provides the climatic data representative of the four ecoclimatic regions along the transect, and Table 2 gives soil classification, site, and vegetation characteristics. The soils included in this study are classified as Black Chernozem, Dark Gray Chernozem, and Gray Luvisol in the Canadian system of soil classification (Agriculture Canada Expert Committee on Soil Survey, 1987). The equivalent classification in *Soil Taxonomy* (Soil Survey Staff, 1975) is Typic Cryoboroll, Argic Cryoboroll, and Cryoboralf, respectively.

All soils are developed on lacustrine clay, with the exception of the soil at site 62O6, which developed on mixed clay and loamy morainal material. Table 3 describes the morphology of each soil along the transect. Table 4 provides additional physical and chemical characteristics for representative Black and Dark Gray soils and two Gray Luvisol soils (one from southern Manitoba and one from the north) which were selected to closely approximate the clay soils encountered in the transect.

Soil Thermal Regime

Soil Temperature

Soil temperatures were measured at 6 depths (5,10,20,50,100, and 150 cm) using thermistors (1986-1989). Sampling frequency averaged 8 times per year in a random manner. Sampling included measurements in January and September of each year, to represent a winter and a late summer measurement, respectively. However, minimum soil temperatures have not been reached at lower depths in January, and, in September, soil temperatures near the surface have started to cool, particularly at the northern sites.

The soil thermal regime at each site was evaluated using the three years of thermistor data from the 20, 50, 100, and 150 cm depths (Table 5). The observed soil temperatures from June 1986 to June 1989 were fitted to an annual sine curve to mathematically calculate the best fitting line to the data and to derive a daily normal temperature for each day of the year (Reimer and Shaykewich, 1980). Then the mean annual soil temperature (MAST) and mean summer soil temperature (MSST) for the period June 1 to September 1 were calculated. Mean January and mean September soil temperature at each depth were calculated from the 1986 to 1989 mid-monthly data.

Soil Thermal Gradients

The latitudinal gradient in soil thermal regime is based on MAST and MSST values calculated from soil temperature measured at 50 cm for the soils at each benchmark site. The data for the nine benchmark sites were grouped and averaged to provide an estimate of the rate of change with increasing northern latitude. Figure 3 includes mean annual air temperatures along the transect. Figure 4 shows the vertical gradient for 1988 for each benchmark site as representative of a winter (January) and late summer (September) regime.

Results and Discussion

Soil Morphology and Properties

Morphological differences among the soils in the transect reflect prevailing climate and vegetation at each site. Black Chernozem soils in the south (sites 62H13 and 62H4) are characterized by dark colored organic-rich Ah horizons (Table 3) due to relatively high levels of organic carbon (Soil 1, Table 4).

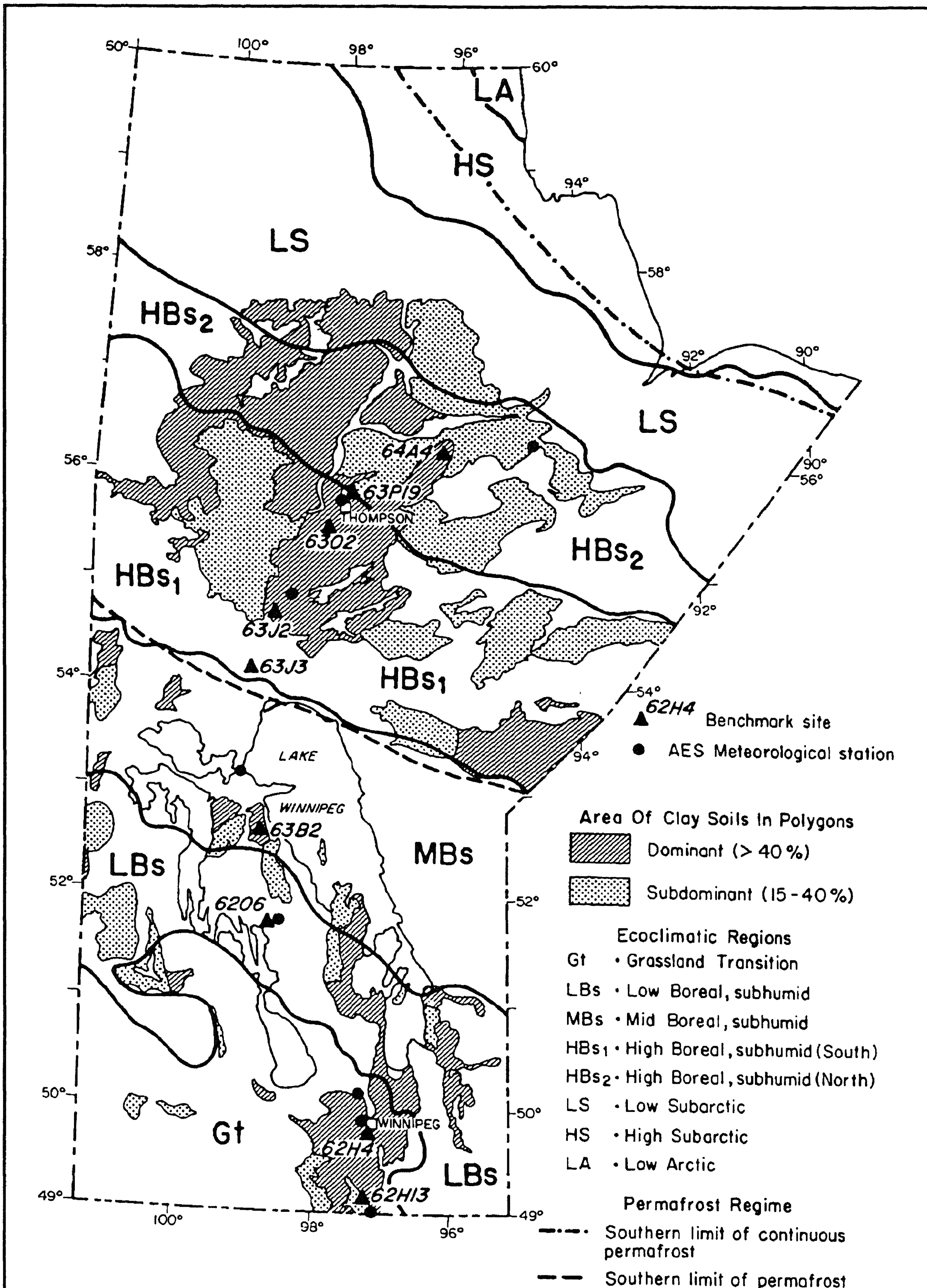


Figure 2. Ecoclimatic regions, permafrost regime, and location of benchmark sites.

The Dark Gray soil (site 62O6) occurs in the transitional zone of mixed grassland and forest vegetation between the Black soils and the Gray Luvisols to the north. Dark Gray soils are characterized by dark colored surface horizons high in organic carbon (Soil 2, Table 4) but which tend to dry to lighter colors and show structural modification due to increased eluviation. The underlying B horizon contains accumulation of clay resulting in coarse granular to fine subangular blocky structure.

Gray Luvisol soils (sites 63B2, 63J3, 63J2, 63O2, 63P19 and 64A4) from the Mid and High Boreal Ecoclimatic regions show stronger leaching as a result of the growth and decomposition of forest vegetation. Their main characteristics

Table 2. Ecoclimatic Region and Benchmark Site Characteristics

Ecoclimatic Regions		Benchmark Site				
Soil Order or Great Group	Vegetation Type (natural)	Site	Soil Subgroup* No.	Elev. (m)	Vegetation Type (actual)	Regional Occurrence of Vertic Characteristics**
Grassland transition region, Gt						
Black Chernozem	Grassland-Aspen Parkland	62H13	Rego Black	234	Oak, aspen	Surface cracking to 1.5 m; slickensides
		62H14	Orthic Black	227	Oak, aspen	
Low Boreal subhumid region, LBs						
Dark Gray Chernozem, Organic	Mixed Deciduous-Coniferous Forest	62O6	Dark Gray	246	Aspen, oak, white spruce	Surface cracking to 0.5 m; slickensides
Mid Boreal subhumid region, MBs						
Gray Luvisol, Organic	Mixed Forest	63B2	Orthic Gray Luvisol	248	Mixed white spruce-aspen	Shallow cracks upper solum; slickensides
High Boreal subhumid region, HBs1						
Gray Luvisol, Brunisolic, Organic	Coniferous and Mixed Forest	63J3	Orthic Gray Luvisol	237	Black spruce, aspen, black poplar, balsam fir	Shallow cracks upper solum
		63J2	Orthic Gray Luvisol	225	Black spruce, black poplar	
		63O2	Orthic Gray Luvisol	220	Black spruce, jack pine	
High Boreal subhumid region, HBs2						
Gray Luvisol, Brunisolic, Organic, Cryosolic	Coniferous Forest	63P19	Solonetzic Gray Luvisol	210	Jack pine, black spruce aspen	Cracking in upper solum; stress
		64A4	Orthic Gray Luvisol (cryic)	180	Black spruce, few white spruce	Surface cracking and slickensides not
observed						
Low Subarctic region, LS						
Brunisolic, Cryosolic	Open Coniferous Forest			No Sites		
High Subarctic region, HS						
Cryosolic, Brunisolic	Open Coniferous Forest & Tundra (shrub)			No Sites		
Low Arctic region, LA						
Cryosolic	Tundra (shrub)			No Sites		

*Soil Classification according to The Canadian System of Soil Classification. Agric. Canada Expert Committee on Soil Survey, 1987.
**Vertic characteristics observed on a regional basis and not necessarily at the benchmark soil site in the transect.

are A horizons with platy or granular structure, which are light colored on drying due to the eluviation of clay, organic matter and iron and aluminum sesquioxides. The underlying illuvial B horizons are characterized by subangular blocky structure (Table 3) and accumulation of clay (Table 4).

Leaching of CaCO3 and reduced pH in the sola accompanied by downward movement of clay and other associated colloidal materials is characteristic of Luvisolic soils (Soil 3 and 4, Table 4). The clayey parent materials of these soils are moderately to strongly calcareous with clay content usually in excess of 60 percent (Table 4). The six Luvisolic soils in the transect occur over a latitudinal distance that results in decreasing soil temperature from south to north and the occurrence of permafrost in the most northerly Luvisol (site 63A4).

The soils in this transect traverse two distinct regions of clay mineralogy. The fine clay fraction in the parent materials of both regions is characterized by mixed mineralogy dominated by illite north of Lake Winnipeg and smectite codominant with illite to the south. Shrink-swell phenomena and development of Vertisols have been observed in such materials but are more common in the south.

Although continuous cracking and slickensides were not noted in the benchmark soils in the transect, observations on similar clayey materials and landscapes in the same regions indicate that these vertic features do occur in Black and Dark Gray soils and occasionally in Gray Luvisols. Cracking between 1 and 1.5 m deep has been observed in Black Chernozems in most years. The cracks are less continuous and more shallow (about 0.5 m in depth) in Dark Gray soils and generally are restricted to shallow cracking in the upper solum of Gray Luvisols (0.3 to 0.5 m).

Slickensides are most common in well and imperfectly drained Black and Dark Gray soils. Prominent slickensides are described in clayey Luvisols in east-central Saskatchewan (Acton et al., 1989) but have not been observed in the colder Luvisols in Manitoba north of Lake Winnipeg. The potential for differential swelling pressures and slickenside formation is reduced in these soils due to cold (cryic) and very cold (pergelic) thermal regimes and the dominance of illitic clay mineralogy in the parent material.

Table 3. Morphological description of benchmark soils.

Horizon	Depth (cm)	Colour (moist)	Texture	Structure	Special Features
	Site 62H13 Rego Black* (Typic Cryoboroll)** 49°08'N, 97°15'W				
L-H	3-0	-	-	-	-
Ah1	0-16	10YR2/1	C	gr	weak tonguing of
Ah2	16-36	10YR2/1	C	gr	A horizons through
AC	36-50	5Y2.5/1	C	gr	AC into Ck
Ck	50-65	2.5Y4/2	SiC	gr	-
2Ck	65-110	10YR4.5/3	SiCL	gr	-
	Site 62H4 Orthic Black* (Typic Cryoboroll)** 49°47'N, 97°08'W				
L-H	2-0	-	-	-	-
Ah	0-24	10YR2/1	C	gr	-
Bm	24-36	10YR3/1.5	C	sbk	-
BC	36-46	2.5Y3/2	C	gr	-
Ck	46-80	2.5Y4/2	C-SiC	gr	silt content increases
2Ck	80-100	10YR6/3	SiCL	gr	below 1 m
	Site 6206 Orthic Dark Gray* (Argic. Cryoboroll)** 51°44'N, 98°46'W				
L-H	10-0	-	-	-	-
Ahe	0-6	10YR2/2	C	gr	-
Bt	6-14	10YR3.5/2	C	gr	-
BC	14-22	10YR3.5/2	C-SiC	gr	-
2Ck1	22-100	2.5Y6/2	SiL	gr	-
2Ck2	100-120	2.5Y6/2	C-SiL	gr	mixed clay and silty till
	Site 63B2 Orthic Gray Luvisol* (Cryoboralf)** 52°39'N, 98°56'W				
L-H	10-0	-	-	-	-
Ae	0-8	10YR4.5/2	L	plt	-
Bt	8-20	10YR3/3	C	sbk	-
Ck	20-140	10YR5.5/2.5	SiC	gr	mixed clay and silty till
	Site 63J3 Orthic Gray Luvisol* (Cryoboralf)** 54°12'N, 99°11'W				
LFH	10-0	-	-	-	-
Ahe	0-5	10YR2/1.5	L	gr	some tonguing of A Ae
Ae	5-13	10YR6/3	FSL	plt	horizon into B horizon
Bt	13-32	10YR3/2	C	sbk	at 25 cm
BC	32-45	10YR4/3	C	sbk	-
Ck	45-100	10YR5/2	SiC	gr	-
	Site 63J2 Orthic Gray Luvisol* (Cryoboralf)** 54°42'N, 98;W				
LFH	12-0	-	-	-	-
Ae	0-7	10YR6/3	SiL	gr	-
Btnj	7-12	10YR4/4	HC	sbk	weak columns with
Bt	12-23	10YR4/3	HC	sbk	shallow cracking in
Cca	23-30	10YR5/4	SiC	gr	upper solum
Ck1	30-105	10YR4/3	HC	abk	-
Ck2	105-115	2.5Y7/2	HC	abk	silty varves at 1.5 m
	Site 6302 Othric Gray Luvisol* (Cryoboralf)** 55°32'N, 98°03'W				
LFH	6-0	5YR2.5/2	-	-	-
Ae	0-7	10YR4/2	C	gr	-
AB	7-10	10YR5/3	C	sbk	shallow cracking in
Btnj	10-17	10YR4/2.5	HC	sbk	upper solum
Bt	17-30	10YR4/2	HC	gr	-
BC	30-36	10YR5/2	HC	gr	-
Ck	36-90	10YR4.5/3	HC	mss	-
	Site 63P19 Solonetzic Gray Luvisol* (Cryoboralf)** 55°55'N, 97°42'W				
LFH	10-0	-	-	-	-
Ae	0-10	10YR4/3	SiCL	plt	-
AB	10-14	10YR5/3	SiC	plt	pronounced vertical
Btnj	14-24	10YR4/3	SiC	col	cracking in upper solum
Bt	24-50	10YR3/4	HC	gr	-
BC	50-60	10YR4/3	SiC	gr	-
Ck1	60-100	10YR3/3	HC	mss	-
Ck2	100-120	7.5YR5/4	HC	mss	-
	Site 64A4 Orthic Gray Luvisol* (Cryoboralf)** 56°17'N, 96°03'W cryic phase				
LFH	25-0	-	-	-	-
Ae	0-7	10YR3.5/2	SiC	gr	cracking absent
Bt	7-18	7.5YR3/2	C	gr	cracking absent
Ckgj	18-28	10YR4.5/4	C	mss	weakly cryoturbated
Ckz	28-50	10YR6/3	SiC	mss	vein ice, cryoturbed

Texture: C:Clay; HC:Heavy Clay; SiC:Silty Clay; SiCL:Silty Clay Loam; Sil:Silt Loam
Structure: plt:platy; abk:angular blocky; sbk:subangular blocky; gr:granular; mss:massive; col:columnar
*Soil classification according to System of Soil Classification for Canada, (Agriculture Canada Expert Committee on Soil Survey, 1987).
**Soil classification according to U.S. Soil Taxonomy (Soil Survey Staff, 1975).

Soils in which vertic features are most common are usually affected by pronounced oscillation between the wet and dry moisture state (Wilding and Tessier, 1988). Northern soils in the transect usually do not dry out to the same degree as the southern soils, due to reduced evapotranspiration resulting from shorter growing season, increased proportion of coniferous vegetation, and lower air temperature causing reduced demand for soil moisture. As a result, soil moisture fluctuation in the clay Luvisol soils, particularly deeper in the profile, is less pronounced than that found in the Black and Dark Gray soils in the south. In addition, in the most northerly soils containing permafrost, gradual melting of the upper layer of frozen material during the thaw season helps to maintain soil moisture levels in the surface soil. Potential for physical movement in soils north of Lake Winnipeg is further reduced by a lower content of high shrink-swell clay, due to the dominance of illitic clay.

Thermal Regime of Clay Soils

The thermal regime of well drained clay soils under forest vegetation shows regional variability along the south-north transect and seasonal variation within the soil profile at each of the nine benchmark sites.

Regional Thermal Regimes in Manitoba

Mean Annual Soil Temperature (MAST) values at 50 cm range from a high of 6.6oC in a Black Chernozem soil to a low of 0.2° C for the northern-most Luvisolic soil affected by permafrost (site 64A4, Table 5). Mean Summer Soil Temperature (MSST) values vary from 12.6°C in the south to 1.1°C at the most northerly Luvisol soil (site 64A4). Soil thermal regime of the Chernozemic Black soils in the Grassland transition region and the Dark Gray Chernozem soil in the Low Boreal region is cold (frigid, in U.S. *Soil Taxonomy* criteria for soils with LFH horizons). Thermal regimes characterizing the Luvisolic soils in the Mid and High Boreal region vary from cold (cryic, in *Soil Taxonomy*) at sites

Table 4. Physical and chemical characteristics of soils selected to closely approximate soil conditions in the south-north transect

Horizon	Depth (cm)	Total Sand %	Silt %	Clay %	pH	$CaCO_3$ Equiv. %	Org. C %	C.E.C. m.e. /100 g
	1. Orthic Black (Typic Cryoboroll)							
Ap 1	0-9	30	27	43	6.7	-	4.5	46
Ap2	9-24	22	32	46	6.7	-	4.4	49
BM	24-42	12	26	62	7.2	-	1.9	48
Bmk	42-65	9	29	62	7.6	8.8	1.1	43
BC	65-78	7	31	62	7.7	14.1	1.0	41
Ck1	78-95	4	31	65	7.8	10.6	-	40
Ck2	95-110	1	32	67	7.8	12.8	-	39
	2. Orthic Dark Gray (Argic Cryoboroll)							
LH	4-0	-	-	-	6.1	2.2	29.7	-
Ahe	0-5	19	26	55	6.1	-	3.6	39
Bt	5-20	15	18	67	7.1	1.6	1.4	37
BC	20-41	15	20	65	7.9	15.3	0.8	24
Ck1	41-76	4	15	81	8.1	16.4	0.1	22
Ck2	76-91	2	8	90	8.1	4.2	-	24
	3. Solonetzic Gray Luvisol-south Cryoboralf)							
LF	5-0	-	-	-	5.0	-	39.8	121
Ae	0-5	9	30	61	5.1	-	2.6	40
AB	5-10	6	29	65	4.7	-	1.3	38
Btnj	10-25	4	21	75	5.0	-	0.8	40
Bt	25-45	1	14	85	6.1	-	0.8	46
BC	45-60	0	11	89	7.5	6.8	0.4	34
Ck1	60-100	1	12	87	7.6	9.2	-	32
Ck2	100-130	1	21	78	7.5	6.0	-	28
	4. Solonetzic Gray Luvisol-north (Cryoboralf)							
LH	2-0	-	-	-	5.9	-	32.4	87
Ae	0-6	3	39	58	5.5	-	4.0	38
AB	6-16	1	38	61	5.6	-	2.5	31
Btnj	16-34	1	46	53	5.9	-	1.0	35
Bt	34-50	0	54	46	7.0	-	0.5	34
Ck1	50-98	0	40	60	7.6	8.3	-	31
Ck2	98-138	0	31	69	7.7	7.7	-	31

63B2, 63J3, and 63J2 to very cold or cryic at sites 63O2 and 63P19. A very cold (pergelic in *Soil Taxonomy* criteria) thermal regime occurs at 150 cm at site 64A4, where the Luvisol soil contains permafrost.

The difference between MAST and MSST decreases with increasing latitude (Figure 3). Differences at 50 cm are greatest in the Black soils in the south (5.5°C at site 62H13 and 6.0°C at site 62H4) and least in the northern Luvisols (0.9°C at site 64A4) affected by the permafrost. These differences are somewhat smaller at the 150 cm depth, ranging from 1.4 to 2.6°C in the 4 southern sites to no difference in northern soils affected by presence of permafrost (Table 5).

As mean annual air temperature decreases, the difference between soil and air temperature tends to increase (Smith et al., 1964). MAST is generally about 3 to 4°C warmer than the corresponding air temperature at latitude 49°N, increasing to about 5°C warmer than the air temperature north of latitude 55°N (Figure 3). At these latitudes, summer soil temperatures are appreciably lower than the air temperature (Smith et al., 1964). In this study, the MSST at all sites (Table 5) is lower than summer air temperature recorded at nearby climatological stations (Table 1).

January soil temperatures are above 0°C at 50 cm throughout the Black Chernozemic clays (sites 62H13 and 62H4) and fall below 0°C in the Dark Gray (site 62O6) and Luvisolic soils (Table 5). Minimum temperature in January at 150 cm remains above 0°C at all sites except in the most northerly Luvisol (64A4). This soil has permafrost at 1.5 m and has a very cold (pergelic) thermal regime. The temperature of the permafrost at this latitude of northern Manitoba remains relatively constant at only a few tenths of a degree below 0°C. The maximum thickness of permafrost at this latitude has been reported to be about 15 m (Brown, 1970), but is much thinner at the benchmark site.

Latitudinal Gradient

Figure 3 shows temperature at 50 cm and the mean annual air temperature along the transect. The regression equations for MAST AND MSST at the 50 cm depth and the mean annual air temperature along the transect show that the rate of temperature change per degree of latitude ranges from a minimum of 0.9°C for MAST to a maximum of 1.5°C for MSST. These estimated rates of decrease in MAST and MSST follow the northward trend of mean annual air

Table 5. Mean soil temperatures at benchmark sites.

Site No.	Depth (cm)	Mean Soil Temperature* Annual	Summer	n	Mean Soil Temperature** January	September
62H13	20	5.7	13.3	23	-1.7	12.5
49°08'N	50	6.0	11.5		0.3	12.2
97°15'W	100	6.0	9.0		2.4	11.2
	150	6.0	7.4		3.9	10.1
62H4	20	4.1	13.1	24	-	11.5
49°47'N	50	6.6	12.6		0.7	13.6
97°08'W	100	6.2	9.5		2.3	12.2
	150	6.3	7.9		4.5	11.4
62O6	20	4.8	10.9	19	-1.7	10.8
51°44'N	50	4.6	9.5		-0.3	10.4
98°46'W	100	4.3	7.6		1.2	9.4
	150	4.6	6.9		2.7	9.5
63B2	20	3.8	9.1	19	-2.0	8.8
52°39'N	50	3.8	7.8		-0.3	8.6
98°56'W	100	4.6	8.5		0.8	8.0
	150	4.2	6.8		1.7	7.4
63J3	20	2.7	7.3	18	-2.3	8.2
54°12'N	50	2.3	5.8		-0.9	7.4
99°11'W	100	2.8	5.0		0.7	7.5
	150	2.7	4.1		1.4	6.6
63J2	20	2.3	7.2	18	-3.3	7.9
54°42'N	50	2.1	5.0		-1.3	7.6
98°58'W	100	1.8	3.0		0.3	5.9
	150	1.7	2.2		0.9	4.9
62O2	20	1.6	5.1	29	-2.4	7.0
55°32'N	50	1.1	3.2		-1.3	5.5
98°03'W	100	1.3	2.4		-0.2	4.9
	150	-0.7	0.9		-0.7***	3.2
63P19	20	1.1	5.0	40	-2.8	6.3
55°55'N	50	1.1	3.4		-0.7	5.3
97°42'W	100	1.1	1.8		0.4	3.8
	150	0.9	1.0		0.6	2.3
64A4	20	0.6	3.7	15	-3.4	5.1
56°17'N	50	0.2	1.1		-1.0	1.9
96°09'W	100	0.1	0.1		-0.0	0.5
	150	-0.1	-0.1		-0.1	-0.0

*Mean soil temperature calculated from thermistor observations, 1986-1989. n=number of observations at each site.
**Mean January and September soil temperature 1986-89, 3 observations.
***January temperature derived from one observation due to sensor failure.

temperature which decreases at a rate of about 1°C per degree of latitude.

Vertical Thermal Regime

Vertical gradients in soil temperature vary with season and latitude. The soil thermal regimes for January and September 1988 are plotted as a function of depth for each site in the transect (Figure 4). These soil temperature gradients are indicative of the maximum seasonal variation at each site. As would be expected, the vertical temperature gradient to the surface is negative in January and positive in September. The mean annual soil thermal gradients (1986-1989 data) also plotted for each site in the transect are nearly isothermal and shift toward 0°C with increasing latitude.

The difference between September and January temperature decreases with increasing latitude as well as with increasing depth in the soil. Greatest seasonal temperature range in the upper 50 cm occurs in the Black soils at the two southernmost sites (62H13 and 62H4) and decreases only slightly to the north along the transect. The range in seasonal temperature variation below 50 cm decreases more rapidly with increasing latitude.

The seasonal temperature difference is least at the northern-most Gray Luvisol (site 64A4). This soil has permafrost at lower depths in the profile and in some summers remains frozen between 50 and 100 cm. The temperature of permafrost at this latitude is only slightly below freezing (Brown 1970) and is virtually isothermal with depth. The occurrence of the permafrost acts as a heat sink and provides a buffer against any large variation in soil temperature near the frost table.

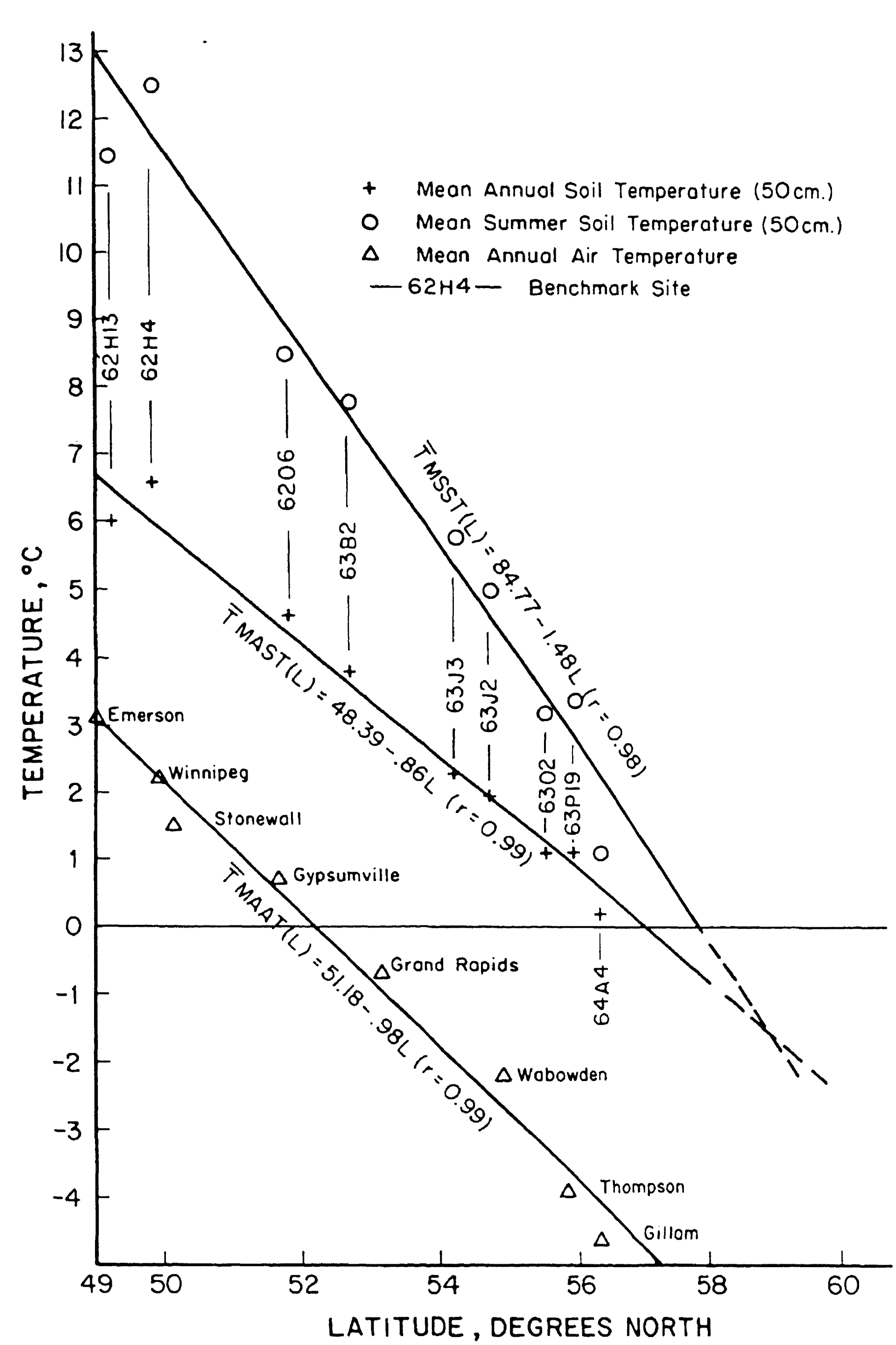


Figure 3. Scatter diagram and fitted lines of mean soil temperature at 50 cm and mean air temperature at 1.2 m.

Summary

Clay soils cover some 7 000 kha in Manitoba between latitude 49°N and 58°N. Soil development and soil thermal regime were examined in relation to climate and vegetation at nine benchmark sites currently under forest in a south-north transect extending from latitude 49°N to 56°17'N. Warmer climate and subhu-

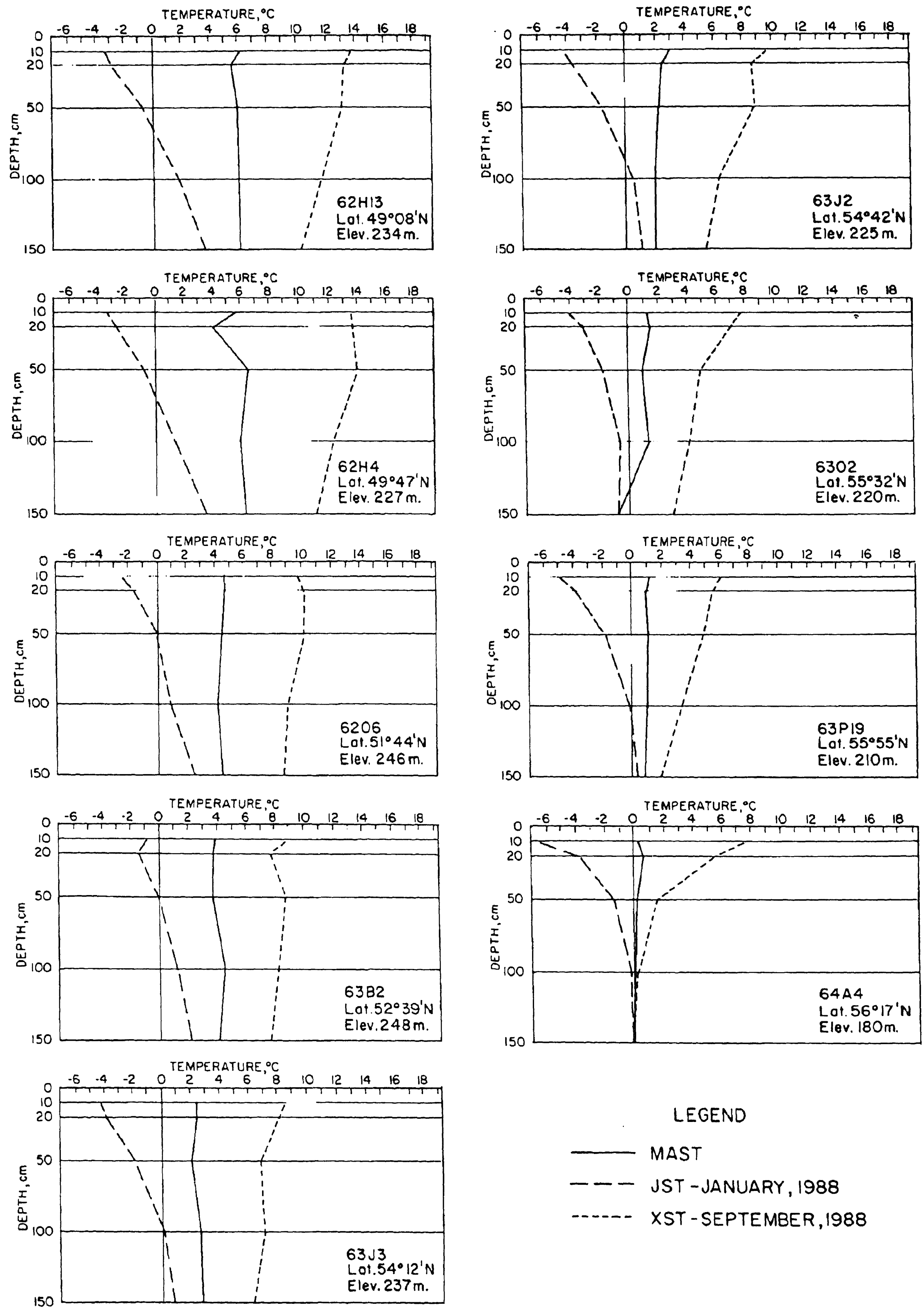


Figure 4. Vertical thermal gradients along south-north transect.

mid conditions in the south result in grassland and grassland-forest transition vegetation characterized by Black and Dark Gray Chernozem soils. Cooler, more humid conditions associated with the Boreal forest extend northward from the grasslands to the southern edge of the Subarctic forest. Soil development in this region results in Gray Luvisol soils; the more severe climate associated with the northern-most Luvisols results in the occurrence of permafrost between 1 and 2 m and cryoturbation of the ground surface. Although vertic soil features were not noted in the soils of the transect, they have been observed in similar clayey materials in the regions covered by this study.

The thermal regimes of clay soils vary from cold (MAST 2-8°C, MSST 8-15°C) for Black and Dark Gray Chernozems to cold and very cold (MAST -7 to 2°C, MSST 5 to 8°C) for the Luvisols. The most northerly clay soils have permafrost within 1.5 m, resulting in a very cold thermal regime at this depth. The calculated rate of decrease in MAST (0.9°C/degree of latitude) and MSST (1.5°C/degree of latitude) shows a northward trend following the mean annual air temperature, which decreases at a rate of about 1.0°C/degree of latitude.

Based on clay content and mineralogy, the soils in the transect have the potential to develop vertic properties (continuous cracking and slickensides). Vertic properties are common in Black and Dark Gray soils which are subject to relatively long periods of drying to depths between 1 and 2 m. These soils have a frigid thermal regime according to U.S. *Soil Taxonomy* criteria and could be classified as Vertisols if the temperature limit is altered to include colder soils.

Although vertic soil properties have been observed in warmer Luvisols, they are not common in clayey Luvisols and Cryosols with colder cryic and pergelic thermal regimes. The colder thermal regime and the occurrence of permafrost in the most northern Luvisols combine with a significantly higher micaceous clay mineralogy, more humid soil moisture regimes, and cryoturbation effects to reduce the potential for development of vertic soil features. As a result, vertic properties are not sufficiently common or well developed in northern Luvisol and Cryosol soils in Manitoba to warrant classification in the proposed Cryert suborder.

Acknowledgements

The authors acknowledge the assistance and cooperation of the following people: Mr. C. Aglugub and Mr. N. Lindberg for data compilation and analysis; Mr. J. Griffiths for technical assistance in preparation of graphics; Mrs. D. Sandberg for typing the manuscript; and staff of the Department of Soil Science, University of Manitoba, and the Canada-Manitoba Soil Survey for consultation and review of the manuscript

References

Acton, D.F., G. Coen and R.E. Smith, 1989. Distribution and properties of clay soils of subarid to subhumid regions of the Interior Plains in western Canada. In Proceedings of the Sixth International Soil Correlation Meeting (ISCOM VI) Saskatchewan, Canada, Montana, Idaho and Wyoming, U.S.A. In Press.

Agriculture Canada Expert Committee on Soil Survey.1987. The Canadian System of soil classification, 2nd ed. Agric. Can. Publ. 1646, 164 pp.

Atmospheric Environment Service. 1980.Canadian Climate Normals. 1957-1980. Vol. 2, Temperature; Vol. 3, Precipitation; Vol 6, Frost.

Bostock, H.S. 1970. Physiographic subdivisions of Canada. pp. 9-30 in Geology and Economic Minerals of Canada, Dept. of Energy, Mines and Resources Canada. Geological Survey of Canada, Economic Geology Report No. 1.

Brown, R.J.E. 1970. Permafrost in Canada. University of Toronto Press. 234 pp.

Brown, R.J.E. and W.O. Kupsch, 1974. Permafrost Terminology. National Research Council of Canada, Technical Memorandum No. 111, 62 pp.

Canada-Manitoba Soil Survey. 1989. Soil Landscapes of Canada-Manitoba, Land Resource Research Centre, Res. Br., Agriculture Canada, Ottawa, Ontario. Agric. Can. Publ. 5242/B (LRRC Cont. No. 87-16)Report and map.

Ecoregion Working Group. 1989. Ecoclimatic regions of Canada, first approximation. Ecoregion Working Group of the Canada Committee on Ecological Land Classification, Ecological Land Classification Series 23. Sustainable Development Branch, Canadian Wildlife Service, Conservation and Protection, Environment Canada, Ottawa, Canada. 119 pp. and map.

Madden, J.G., personal communication. Clay mineralogy of some Manitoba parent materials. Canada-Manitoba Soil Survey, 362 Ellis Building, University of Manitoba, Winnipeg, Manitoba. R3T 2N2.

Manitoba Ecoclimatic Region Working Group. 1985. Ecoclimatic Regions of Manitoba, Canada-Manitoba Soil Survey, 362 Ellis Building, University of Manitoba, Winnipeg, Manitoba. R3T 2N2.

Prest, V.K. 1970. Quaternary Geology of Canada. pp.77-764 in Geology and Economic Minerals of Canada, Dept. of Energy, Mines and Resources, Canada. Geologiucal Survey of Canada, Economic Geology Report No.1.

Reimer, A. and C.F. Shaykewich. 1980. Estimation of Manitoba soil temperatures from atmospheric meteorological measurements. Can. J. Soil Sci. 60:299-309.

Smith, G. D., L.H. Robinson, and D. Swanson, 1964. Soil temperature regimes-their characteristics and predictability. SCS-TP-144 Soil Conserv. Serv., U.S. Dep. Agr., Washington, D.C.

Soil Survey Staff. 1975. Soil Taxonomy. Soil Conservation Service. U.S. Department of Agriculture. Agricultureal Handbook No. 436. Washington D.C., 754 pp.

Wilding, L.P. and D. Tessier. 1988. Genesis of Vertisols: Shrink-Swell Phenomena. pp.55. In L.P. Wilding and R. Puentes (eds). Vertisols: Their Distribution, properties classification and management, Texas A & M Press. College Station.

Periglacial Features as Sources of Variability in Wyoming Aridisols

L.C. Munn[1]

Abstract

Periglacial features including ice wedge casts, ground wedges and low-relief soil mounds are common in the basins of Wyoming. These features represent former permafrost environments, during glacial episodes. Soil formed on this patterned ground are younger and show different morphology and chemistry compared to older, undisturbed soils on the same landscape. The features thus contribute to soil variability; this variability influences present-day vegetation distribution and yield.

Introduction

The high, nonglaciated basins of Wyoming contain many relicts of previous periglacial environments—casts of ice wedges, ground wedges, and mounded topography (Mears, 1987; Nisson, 1985; Spackman and Munn, 1984). In the surrounding mountains, glacial tills and sorted patterned ground features attest to the extent of the glacial climates (Richmond and Fullerton, 1986). The purpose of this paper is to discuss the contribution of these relict features to the spatial variability in morphology and chemistry on Aridisol landscapes.

Methods

Data for this paper are taken largely from previously published work by the author, from Wyoming soil survey reports, or from data collected by the National Soil Survey Laboratory (1988). Methods of data collection and laboratory analyses are given in the documents cited and generally represent standard methods specified in *Soil Taxonomy* (Soil Survey Staff, 1975) and the *Soil Survey Manual* (Soil Survey Staff, 1981).

Periglacial Features: Mode of Formation

Black (1976) details the formation of both ice wedges and ground wedges in modern permafrost environments. Mears (1981, 1987) has documented multiple occurrances of both features in Wyoming's cold basins. Both types begin with thermal cracking-contraction cracks of 1 to 5 mm wide that form during the cold period of the annual climatic cycle. In moist permafrost environments, water from the thawing active layer fills the cracks in the spring. In dry permafrost environments, blowing soil particles fill the cracks during the cold period. In both cases, the initial filled cracks in the permafrost layer act as a "memory" and subsequent cracks open each successive season in nearly the same location until ice or ground wedges of up to several meters in cross section have built up (Figure 1). In three dimensions, the cracks form polygons with diameters of 1 to 40 m.

Washburn (1980) proposed that a mean annual air temperature of -5°C was required to form wedges. If this threshold temperature is valid, then mean annual temperature in Wyoming during the Pleistocene must have veen at least 10°C (Laramie Basin) to 13°C (Bighorn Basin) colder than the present climate. In addition to a subzero mean annual temperature, a rapid drop in temperature, variously estimated at from 4°C to 10°C (Washburn, 1980), is required to initiate cracking of the frozen soil, sediment, or rock. The presence of frozen water is required for the formation of continuous cracks in loose soils or sediments. Walker (1987) describes a fossil animal assemblege for northwestern Wyoming that agrees with the concept of a permafrost tundra.

Another feature, thought to be an indicator of a periglacial environment, is large (8-10 m diameter), low relief (0.5-1 m) mounds. These mounds are spaced regularly across large areas of Pleistocene-aged surfaces in Wyoming (Spackman and Munn, 1984). Spackman and Munn (1984) hypothesized that the mounds were formed by cryoturbation as a result of pressure engendered by water being trapped between a downward freezing front at the top of the active layer and an upward freezing zone from the top of the permafrost. In coarse-textured materials, water migration to the freezing

[1]Department of Plant, Soil and Insect Sciences, University of Wyoming, Laramie, WY 82071.

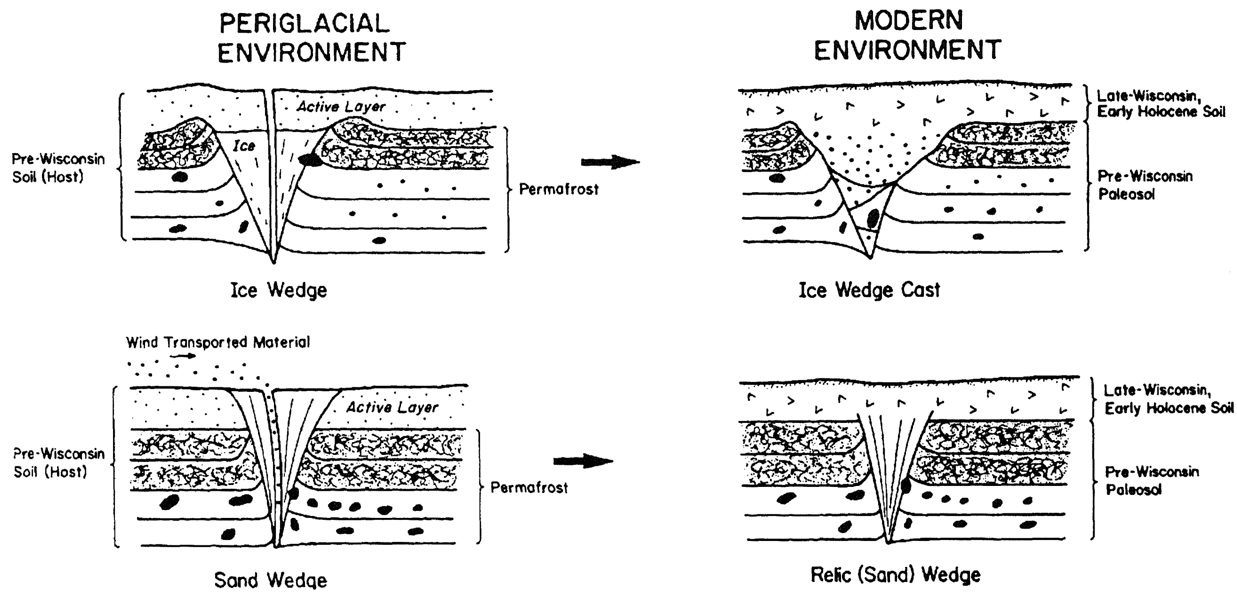


Figure 1. Schematic representation of mode of formation of ice and sand (ground) wedges (Munn, 1987).

fronts might be slow enough for this to occur during rapid freezing.

Another hypothesis is that the mounds represent "mudboils" comparable to those in the modern arctic on materials with low plastic and liquid limits (Shilts, 1978). Mudboils are a form of nonsorted circles (Washburn, 1980) that form during the annual thawing cycle of the active layer. Positive pore water pressure is developed because the permafrost layer prevents drainage of meltwater and supersaturation results as ice lenses melt. Pressure built up under a confining surface layer, or carapace (in this case, the Haplargid Bt horizon), is released through an upwelling of a soil and water slurry. This pressure release mechanism accounts for the statistically regular distribution of the mounds (Spackman and Munn, 1984).

Discussion

Both types of wedge features and the cryoturbation mounds (Figure 2) are important sources of variability on Pleistocene-age landscapes. Table 1 summarizes variations in textural properties, morphology, and carbonate accumulations for several sites in Wyoming. Note that the wedge or mound soils are typically in a different (coarser) textural family than the host soil. The mounded landscapes are typically Camborthid (mound)/Haplargid (intermound) complexes. In the Laramie Basin, the mound soil provides a sandy range site, while the intermound soil is in a loamy range site. Depth of penetration of precipitation is typically greater in the coarser-textured wedge or mound soils.

Table 1: Comparison of Texture and Related Properties in Former-Permafrost Landscapes, Wyoming

Feature	Percent of Landscape	Soil Family	Maximum Percentage Clay	$CaCO_3$	Gravel in B Horizon % by volume	Electrical Conductivity at 100 cm depth ($ds \cdot m^{-1}$)	Carbonate Accumulation[1] ($kg \cdot m^{-2}$)
Ice Wedge Casts Rawlins Area (Munn, 1987)	30 (Wedge)	(W)[2] Typic Haplargids, coarse-loamy, mixed, frigid.	16	9	5	1.9	92
		(H) Typic Haplargids, fine-loamy, mixed, frigid.	31	45	25	>15.0	250
Ground Wedges, Laramie Area (Munn and Spackman, in review)	70 (Wedge)	(W) Borollic Natrargids, fine-loamy, mixed.	33	7	<1	8.5	171
		(H) Borollic Natrargids, fine, mixed.	39	10	3-5	12.3	419
Ice Wedge Casts, Laramie Area (Spackman and Munn, 1984)	35 (Wedge)	(W) Borollic Haplargids, coarse-loamy, mixed.	15	11	6	0.6	113
		(H) Borollic Haplargids, fine-loamy, mixed.	33	25	20	0.7	250
Cryoturbation Mound, (Spackman, 1982)	35 (Mound)	(M) Borollic Camborthids, coarse-loamy, mixed.	10	7	32	10.9	169[3]
		(H) Borollic Haplargids, fine-loamy, mixed.	33	25	20	0.7	250
Ground Wedge, Kemmerer Area (National Soil Survey Laboratory, 1988)	5-10 (Wedge)	(W) Typic Haplargid, fine-loamy, mixed, frigid.	24	18	TR	3.3	247
		(H) Borollic Haplargid, fine, mixed.	47	29	1	4.3	373

[1]Carbonate accumulation to varying depths to include entire solum.
[2]W = Wedge, H = Host, M = Mound.
[3]Includes primary carbonate from parent material (ruptured host Bk horizons).

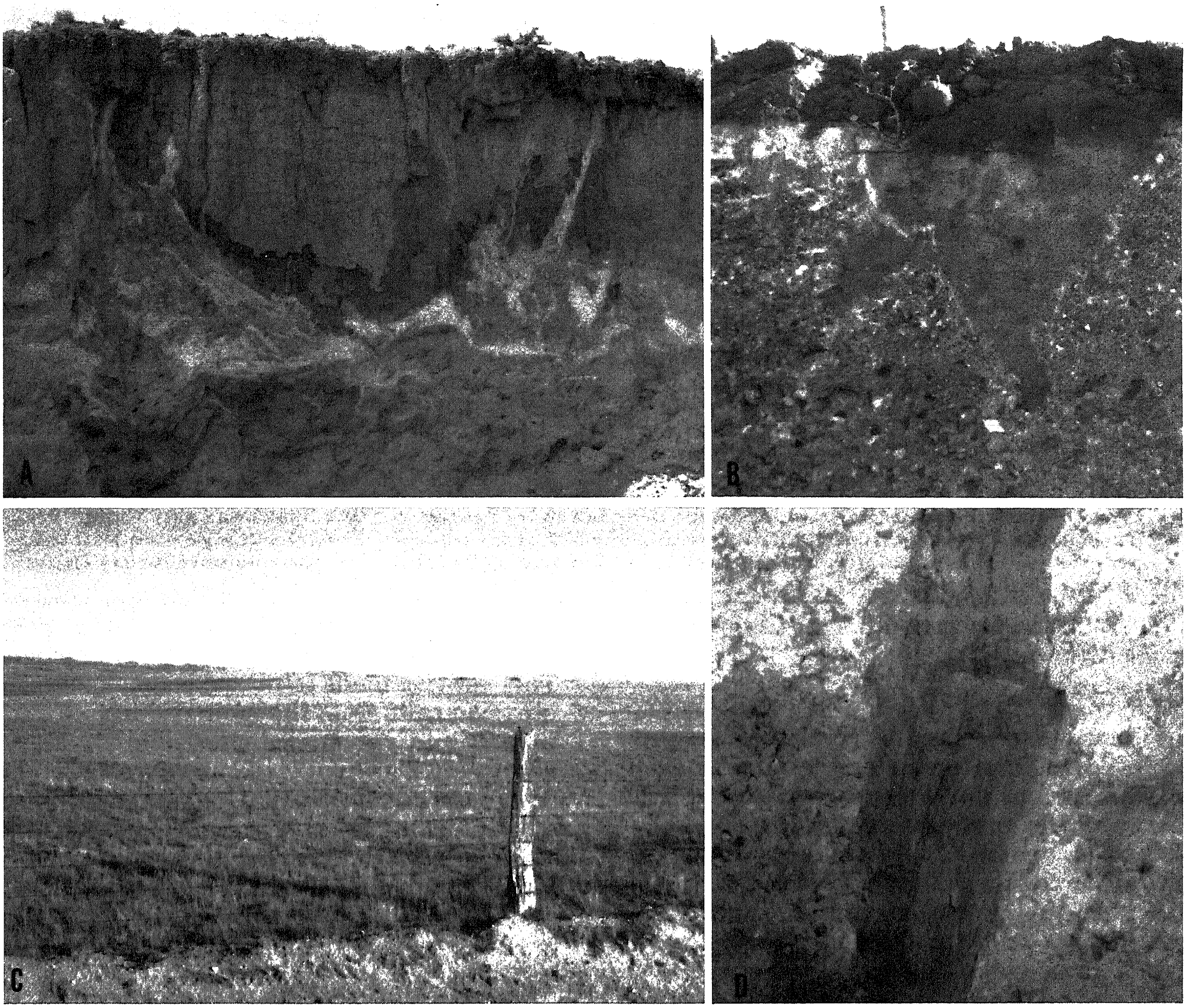


Figure 2. Composite of (a) ground wedges near Laramie (Munn and Spackman, in review), (b) ice wedge cast near Rawlins (Munn, 1987), (c) mounded topography near Laramie (Spackman and Munn, 1984), and (d) ground wedge showing vertical fabric (Munn and Spackman, in review). Both wedge features show contortion of the Bk horizon of the wedge host soil adjacent to the wedge.

Plant Response

Catt (1986) describes crop response to ice wedge casts in Europe. In England, yields of both barley and wheat were approximately twices as high on fine-textured wedge fills as on the coarse-textured polygon interiors (e.g. 5.06 tonne ha^{-1} vs. 2.11 tonne ha^{-1} for wheat). When the wedge fill material was coarser-textured than the material of the wedge host, grain yields were lower on the fills at sites in England, Denmark, and Germany.

In arid climates, higher biomass production typically occurs on coarse-textured soils (and wedge fills) where storage of plant-available water is greater. At the Kemmerer site, plant available water holding capacity in the wedge fill is .31 cm cm^{-1}, compared to .20 cm cm^{-1} in the host soil (Natinal Soil Survey Laboratory, 1988). Water held at fifteen bars is 9% in the wedge, in the host soil it is 15%. Where the wedges penetrate dense Bk horizons, as at this site, improved grass growth over the wedge fill

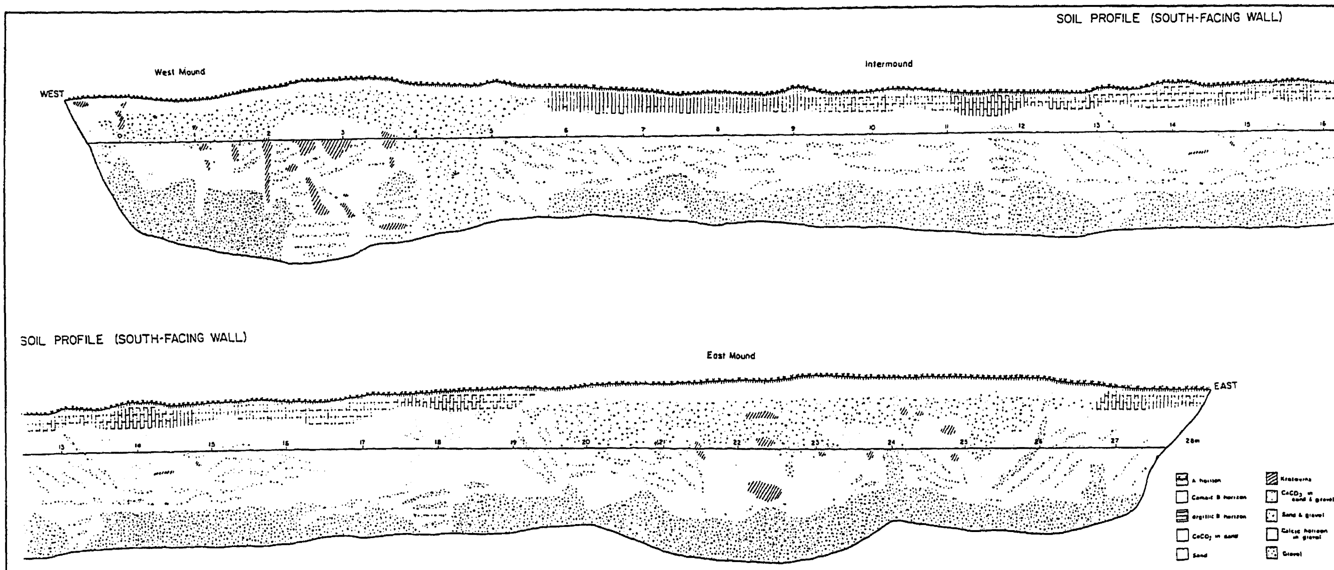


Figure 3. Cross section of mounded landscape near Laramie (Spackman, 1982; Spackman and Munn, 1984). The mounds (Camborthids) occupied approximately 35% of the transect. Fossil ice wedge casts were present at several places along the transect, e.g., 11 to 12 m and at 13 m and 18 m. Intermound soils are Haplargids with calcic horizons.

often results in a clear outline of the polygonal wedge pattern.

Soil Morphology and Genesis

Soil mounds produced by cryoturbation show a definite admixture of materials from pre-existing soil horizons, including carbonates and soluble salts. In contrast, material in ice wedge casts originated in the active layer (annual thaw zone above the permafrost). Material from this layer slumped into the depression formed as the ice wedge melted with the final thawing of the permafrost. For ground wedges, the wedge fill originated as aeolian material locally derived from the surface of the active layer.

Soils developed on the mounds contain diffuse "primary" carbonates inherited from the parent material, in addition to secondary carbonates, which occur as threads, rock-rinds, and nodules. Carbonates in ice wedge casts and in the ground wedges represent "secondary" carbonates. In both mound and wedge soils, the accumulations of clay, carbonate, and gypsum are less than in the host soils, as would be predicted by their younger age. Rates of accumulation of clay and carbonates seem to be very similar to those described for the Las Cruces, New Mexico area by Gile and Grossman (1979).

The Laramie Basin ground wedges and the ground wedges at Kemmerer are interpreted as being of Bull Lake (Illinoian) age, while the other features are thought to be of Pinedale (Wisconsin) age. Rodents (*Cynomys spp, Spermophilus richardsmi*) prefer the mounds for burrow sites, and excavations of the mounds reveal active burrows as well as krotovinas. This activity tends to retard leaching of carbonates and salts from the mounds, despite coarser texture and gravel content (Table 1). Figure 3 illustrates the complexity produced on a landscape with wedge features, mounds, and rodent activity in the Laramie area (Spackman, 1982). The mound-forming event disrupted the original loamy over gravelly sand stratification in the alluvial parent material, mixing gravel throughout the profile of the mound soil.

Conclusions

The ubiquitousness of the wedges and mounds in Wyoming's high basins (Mears, 1987; Nisson, 1985) and the marked differences in soil morphology and chemistry between host and wedge or mound soils make their features important contributors to the complexity of Wyoming's desert landscapes. Most stable, older-than-Holocene surfaces in western Wyoming's basins have polygenetic soils related to former permafrost climates.

Literature Cited

Black, R.F. 1976. Periglacial features indicative of permafrost: ice and soil wedges. Quat. Res. 6:3-26.

Catt, J.A. 1986. Effects of the Quaternary on soils of present land surfaces. p. 206-232. In Soils and Quaternary Geology. Clarendon, Oxford.

Gile, L. H. and R. B. Grossman. 1979. The desert project soil monograph. USDA. U.S. Gov't. Print. Office, Washington DC.

Mears, B., Jr. 1981. Periglacial wedges and the late Pleistocene environment of Wyoming's intermontane basins. Quat. Research 15:171-198.

Mears, B., Jr. 1987. Late Pleistocene periglacial wedge sites in Wyoming: an illustrated compendium. Memoir No. 3. The Geological Survey of Wyoming, Laramie.

Munn, L.C. 1987. Soil genesis associated with periglacial ice wedge casts, southcentral Wyoming. Soil Sci. Soc. Am. J. 51:1000-1004.

Munn, L.C. and L.K. Spackman. In review. Soil genesis associated with periglacial ground wedges, Laramie Basin, Wyoming. Quaternary Res.

National Soil Survey Laboratory. 1988. Pedon 88P0858 Tresano Varient. USDA- SCS, Lincoln, NE.

Nisson, T.C. 1985. Field and laboratory studies of selected periglacial wedge-polygons in southern Wyoming. M.S. thesis. Univ. of Wyoming, Laramie.

Richmond, G.M., and Fullerton, D.S. 1986. Summation of quaternary glaciations in the United States of America. Quaternary Science Reviews, Vol. 5. Pergamon, New York.

Shilts, W.W. 1978. Nature and genesis of mudboils, central Keewatin, Canada. Can. J. Earth Sci. 15:1053-1068.

Soil Survey Staff. 1975. Soil taxonomy. Agric. Handbk. No. 436. USDA. U.S. Govt. Print. Office, Washington DC.

Soil Survey Staff. 1981. Soil survey manual. Chapter 4. Working Draft (430-V-SSm). U.S. Govt. Print. Office, Washington DC.

Spackman, L.K. 1982. Genesis and morphology of soils associated with formation of Laramie Basin (Mima-like) mounds. M.S. thesis. Dept. of Plant Science. Univ. of Wyoming, Laramie.

Spackman L.K. and L.C. Munn. 1984. Genesis and morphology of soils associated with formation of Laramie Basin (Mima-like) mounds in Wyoming. Soil Sci. Soc. Am. J. 48:1384-1392.

Walker, D.N. 1987. Late Pleistocene/Holocene environmental changes in the northwestern plains of Wyoming: the mammalian record. Illinois State Museum Scientific Paper 22:334-392.

Washburn, A.L. 1980. Geocryology: a survey of periglacial processes and environments. London/Halstead, New York.

A Comparison of Land Use and Productivity of Clay and Loam Soils within the Interior Plains of Western Canada

C. Onofrei[1], J. Dumanski[2], R.G. Eilers[1], and R.E. Smith[1]

Abstract

Clay and loam soils constitute major land resources for crop production in western Canada. In the present study, two separate investigations within the Chernozemic soil zone (Brown, Dark Brown, and Black) were completed. The first was a comparative land use analysis for farms on clay and loam soils in Saskatchewan, based on data from Statistics Canada; the second was a crop production assessment for clay and loam soils in southern Manitoba, using a wheat yield modeling simulation technique.

Spring wheat is the dominant crop grown in each soil zone. Results of the land use analysis showed that farms in the Brown zone operate marginally more efficiently than in the other areas, with slightly better efficiency on clay than on loam soils. Estimates of efficiency were based on ratios of sales to expenses.

The crop productivity analysis indicated that wheat yield frequency distributions varied not only between the two textural groups, but also within the same textural groups within a given soil zone. The results of this analysis also indicated that certain soil characteristics, agronomic land attributes, and weather elements strongly affect productivity in the interior plains of western Canada. Progress in yield prediction research can be made by focussing on these key controling factors.

Introduction

Production of grains and oilseeds is the main agricultural activity in the interior plains of western Canada. Generally these commodities are produced on land resources consisting of loam and clay soils. The term "land" as used in this paper includes elements of soil as well as elements of geology, the atmosphere, land management, and land use (FAO, 1984).

A system's efficiency is determined as the ratio of its performance to the costs involved. The general indicator used to measure the performance of a crop-oriented agricultural system is annual yield. Due to the stochastic character of weather, however, annual crop yields are often highly variable, particularly these obtained within the Canadian Prairie region.

In general terms, the climate of the southern Prairies is characterized by south to north gradients of energy and moisture; the temperature decreases while the moisture increases with increasing latitude. However, crop yield variability on scales of space and time exhibits a more complicated and complex pattern than can be explained by simple correlation of yield with these two factors.

The objectives of this study were: (1) to compare the general socialeconomic characteristics of farms that operate on clay and adjacent loam soils within the three major soil zones, and (2) to illustrate the variability in crop yield characteristic of clay and loam soils within the Black zone of southern Manitoba. These latter characteristics illustrate the relative levels of production risk between these two types of soil as calculated by yield probability functions.

Clay soils occupy a significant area in western Canada although they are less extensive than loam soils. About 5.6 million ha of clay soils are located in the southern Prairies in the Brown (1.2 million ha), Dark Brown (1.9 million ha), and Black (2.5 million ha) soil zones. These zonal Great Groups are distributed approximately concentrically from the south-central portion of the region. Figure 1 schematically presents the location of study sites

Data Used and Methods of Analysis

Land Use Investigation

For the land use investigation two adjacent polygons were selected from each of the three Chernozemic soil zones of southern Saskatchewan. The polygons were selected from the 1:5 M Soils of Canada map (Clayton et al., 1977) and data for these polygons was obtained

[1]Land Resource Research Centre (LRRC), Research Branch, Agriculture Canada, University of Manitoba, Winnipeg, Manitoba.

[2]LRRC, Research Branch, Agriculture Canada, Ottawa, Ontario.

from the Land Potential Data Base (Kirkwood et al., 1989).

Data for the land use investigation were derived from the 1981 Census of Agriculture. Methods employed were descriptive statistics and simple economic analysis on an "average" farm data set for each polygon (Huffman, 1988). The representative data set was obtained by overlaying data from Census Enumeration Area (EA) maps on the Soils of Canada map.

The arithmetic mean was considered an appropriate statistic for describing the central tendency of physical variables and crop distribution characteristics such as farm size and crop areas. Since some farms on the Prairies are mixed farms, the distribution of economic variables can be highly skewed by extremely small or extremely large data. To avoid the effect of such data on the measure of location, the median was selected as the most appropriate statistic for describing economic variables such as capital value, expenses, and sales.

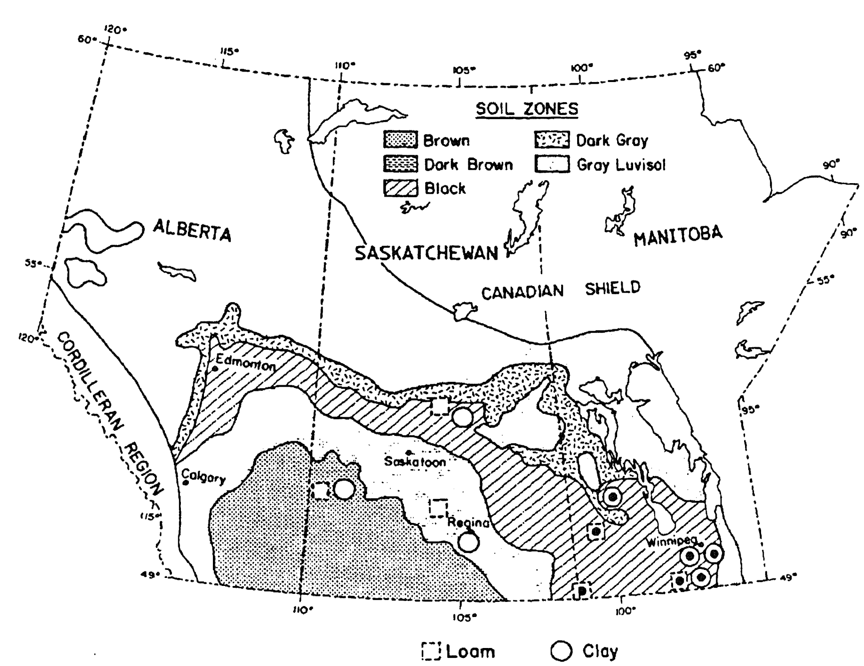


Figure 1. Schematic representation of the location of sites included in the land use investigation (□ ○) and in the crop productivity assessment (▣ ⊙).

Crop Productivity Assessment

Modeling simulation techniques were employed in the crop yield investigation. The study focused on the combined effects of soil properties, management, and climate/weather elements on crop production. Soil series were used as the soil component in the model, but since a soil series may occur in different polygons (geographical locations), crop yields obtained on similar soils in different locations could be quite different.

The modeling procedure was based on a simulation technique developed for land evaluation (Onofrei, 1986). Spring wheat is the most important commodity in the Prairie region, and it was used as the indicator crop. Since yield is controlled primarily by the weather pattern specific to each growing season, PIXMOD, a physical model, was run repeatedly for each polygon/soil, using historic daily weather records for the period 1964 to 1983. Overall, this provided yield frequency distributions, which give insights into natural risk for crop production.

PIXMOD differs from other simulation models in the sense that it operates with readily available data sets (soil, weather, and management) from soil surveys, soil testing laboratories, and climatic stations. It uses potential wheat yield estimates as calculated by the FAO model (FAO, 1978; Dumanski and Stewart, 1981). The potential wheat yield is considered a theoretical limit for photosynthesis, and it is used in this study as the agronomic ceiling.

It should be noted that the term "potential" is used in modeling and in this study in a manner similar to the use of the potential evapotranspiration term in agrometeorology; i.e.,the state of the system in a standard condition. The potential yield is defined as the wheat yield that can be obtained on a given land area if all matter-energy forms required by the crop, with the exception of photosynthetically active radiation, are maintained at optimum over the entire growing season. In real life, such conditions virtually never exist. However, based on potential yield and daily calculation of phenological crop development, soil moisture content, nitrate-nitrogen, and soil temperature, PIXMOD simulated the above-ground dry matter for a growing season, and this was converted into grain yield, using the harvest index approach.

The productivity model was run for seven areas (polygons) in southern Manitoba (Table 1), as identified on the Generalized Soil Landscape map of Manitoba1. These were selected to reflect difference in productivity between loam

and clay soils in adjacent areas. Four polygons have dominantly clay soils, 89-Red River (89-RIV), 90-Osborne (90-OBO), 75-Dencross (75-DCS) and 125-Dauphin (125-DPH), and three polygons have dominantly loam soils, 91-Reinland (73-RLD), 53-Newdale (53-NDL), and 1-Ryerson (1-RYS). All polygons are located in the Black soil zone.

Table 1. Location and general characteristics of soil series included in the crop productivity analysis.

Polygon No.[1]	Coordinate of centroid polygon		Soil series name CanSis code[2]	Profile texture	Drainage class	Water table depth (cm.)	Classification Canada[3]	U.S.A.[4]
	longitude	latitude						
89	97°17'	49°44'	RIV	Clay	Imperfect	> 150	GLRBL	ACB
90	96°51'	49°41'	OBO	Clay	Poor	≤ 120	RHG	TCA
75	97°17'	49°05'	DCS	Clay	Imperfect	> 150	GLRBL	ACB
73	98°00'	49°12'	RLD	Loam	Imperfect	> 150	GLRBL	AHB
125	100°00'	51°08'	DPH	Clay	Imperfect	> 150	GLRBL	ACB
53	100°35'	50°27'	NDL	Loam	Well	> 150	OBL	TCB
1	101°16'	49°03	'RYS	Loam	Well	> 150	OBL	UHB

Note:
[1]Polygon number from GSLM (1:1M);
[2]Soil Series name: RIV = Red River, OBO = Osborne, DCS = Dencross, RLD = Reinland, DPH = Dauphin, NDL = Nwdale, RYS = Ryerson;
[3]Subgroup: GLRBL - Gleyed Rego Black (Chernozem), RHG - Rego Humic Gleysol (Gleysol), OBL - Orthic Black (Chernozem);
[4]U.S.A. Taxonomic equivalent: ACB - Aquic (Vertic) Cryoboroll, TCA - Typic Cryaquoll, AHB - Aquic Haploboroll, TCB - Typic Cryoboroll, UHB - Udic Haploboroll.

Three polygons with clay soils (89-RIV, 90-OBO, and 75-DCS) and one polygon with loam soil (73-RLD) occur in close proximity to each other and thus have similar weather conditions. Therefore, considering that management input was kept constant in each model run, the differences between the yields in these polygons reflect mainly the impact of different soil attributes on productivity. The 125-DPH clay soil and the 53-NDL loam soil are located near the northern limit of the Black Chernozemic zone in Manitoba, while the 1-RYS loam soil is located in the southwest corner of Manitoba which is characteristically the driest area in the province. The Red River (89-RIV) and Dauphin (125-DPH) clay soils are closely related to each other, as are the Newdale (53-NDL) and Ryerson (1-RYS) loam soils, in terms of their major soil characteristics and agronomic attributes. The differences between yields on these soils, therefore, reflect mainly the impact of regional weather on land productivity.

Three categories of data were compiled for the crop production investigation, namely soil, daily weather, and management data. The soil data included: rooting depth, surface texture, % of coarse fragments (gravel), drainage class, infiltration rate, water table depth, incoming runoff, shape and frequencies of unconnected microdepressions, slope class of depressions, and diagnostic soil profile horizons. For each identified master horizon, the centre point, bulk density, particle size distribution (clay, silt, very fine sand, fine sand), organic carbon, volumetric field capacity, volumetric wilting point, and volumetric water content at seeding time also were considered in the model. All soil properties/attributes were derived from the Canadian Soil Information System (CanSIS) and from different soil survey reports.

Weather data included: precipitation, maximum air temperature, minimum air temperature, solar radiation at the top of the atmosphere, and photoperiod. A weather file was created for each of the selected polygons, using the Thiessen weighting procedure (Williams and Hayhoe, 1982) to extrapolate from observed data recorded by the Atmospheric Environment Service (AES).

Of the many management data which influence crop yields, the following were considered the most relevant for modeling wheat production in the Prairie region: seeding date, soil nitrate-nitrogen content at seeding time, nitrogen fertilizer applied (amount and date of application), harvest date, and the agronomic crop yield potential. The latter parameter, although not fully controlled by the management factors, is used as an indication of the agronomic ceiling for a given crop for given management inputs and technology (applied knowledge).

The simulated yield series for each soil in each location (polygon) were analyzed statistically, employing the following tests:

a) the Quade test and the Friedman test (Conover, 1980) to determine if all soils have identical effect on yield
b) the Shapiro-Wilk test of normality for each yield series
c) transformation of nonnormally distributed data series to yield a family of normal distributions
d) calculation of location and scale parameters (μ and σ) for each distribution function.

The details of these statistical analyses are presented in a more comprehensive land evaluation study (Onofrei, 1987).

Results and Discussion

The data presented in this study pertain to the dominant conditions in each polygon. Although local soil properties, weather, or management can have significant impact in a limited area, it is considered that these effects are small overall in comparison to the geographic location of the area. In the same vein, the data pertain to specific polygons, which, although suitable for comparative analysis, may not necessarily describe accurately a larger zone.

Land Use Characteristics

In general, farm size and the proportion of summerfallow land decrease on both clay and loam soils from the Brown to the Black soil zone (Table 2). Summerfallow accounts for about 40% of the use of arable land on farms in the Brown zone and 37% on Dark Brown, but this decreases to about 16-20% in the Black zone. Along with decrease in farm size, there is a progressive increase in absolute value of land and in the marketable production per hectare. These characteristics reflect the fact that large farms tend to concentrate in the Brown and Dark Brown zones where productivity per unit area is low, the risks in crop production are higher, and crop rotations that include fallow once in 2 or 3 years are standard farming practices. However, with decrease in farm size and larger proportion of land used for continuous cropping, as in the Black soil zone, there is a progressive increase in total capital investment and operating expenses.

Farms on clay soils tend to be smaller than those on loam soils in the Brown, but conversely they are larger in the Black zone. The proportion of area cultivated is 5-7% larger on clay soils in each zone. In general there are no major differences between the two textural groups in proportions of field crops grown, but clay soils are used more commonly for cereal crops. On loam soils the proportion of cereal crops decreases in favor of forage crops.

A noticeable difference exists, in terms of economic characteristics, between farms on clay soils and those on loam soils in all zones. For example, total capital investment on clay vs. loam soils is 23% higher on Brown soils, 56% higher on the Dark Brown soils, and 21% higher on the Black soils. Also, the operating expenses on clay soils are 11-28% greater than on the loam soils. Overall, farms on clay soils outperform these on loam soils in terms of total sales per unit area and provide better returns to investment (Huffman, 1988). Based on 1981 data, the value of production per cost of inputs ranged from 2.82 on clay soils in the Brown to 1.99 on clay soils in the Black.

Table 2. Land Use characteristics on clay and loam soils with the Chernozemic Great Groups.

	Brown			Dark Brown			Black		
Characteristic	Clay A1013 - KIN	Loam A1019 - ALS	Clay/ Loam ratio	Clay A2055 - REG	Loam A2039 - DIV	Clay/ Loam ratio	Clay A3098 - MEL	Loam A3086 - PRA	Clay/ Loam ratio
Farm Size (ha)	483.54	516.19	0.94	384.94	397.19	0.97	280.60	251.20	1.12
Cultivated/Farm (%)	93.42	88.67	1.05	94.99	88.78	1.07	90.41	83.18	1.09
Summerfallow (%)	41.50	41.36	1.00	36.26	37.45	0.97	16.07	19.23	0.84
Wheat (%)	44.66	46.25	0.97	50.37	41.31	1.22	33.10	30.26	1.09
Oilseed (%)	0.65	0.28	2.32	0.95	0.97	0.98	13.08	10.59	1.24
Other Grain (%)	9.54	5.11	1.87	4.99	8.13	0.61	28.48	20.26	1.41
Hay (%)	0.54	1.54	0.35	2.63	5.45	0.48	3.29	9.75	0.34
Improved Pasture (%)	1.54	3.91	0.39	2.30	3.55	0.65	2.11	4.44	0.48
Total Capital Investment ($/ha)	1926.72	1563.69	1.23	2094.71	1338.71	1.56	2423.31	2008.88	1.21
Operating Expenses ($/ha)	50.64	45.49	1.11	53.43	46.49	1.15	99.54	77.97	1.28
Total Sales ($/ha)	151.57	129.29	1.17	143.66	121.83	1.18	203.66	155.48	1.31
Sales/Investment	0.08	0.08	1.00	0.07	0.09	0.78	0.09	0.08	1.13
Sales/Expenses	2.82	2.69	1.05	2.56	2.54	1.01	1.99	2.01	0.99

Crop Productivity

Running PIXMOD for all polygons considered in the study, using 20 years of historic weather data, resulted in seven series of annual wheat yield values. Assuming normal distribution of wheat yield on every soil/polygon included in the analysis, and pooling the yields on clay soils and the yields on loam soils, the overall average yield was 2025 kg/ha on clay and 2046 kg/ha on loam. This might suggest at first glance that wheat yields are about uniform across the Black Chernozem soil zone in Manitoba, with no significant differences between yields obtained on clay soils and those obtained on loam soils. In reality, this is not the case.

Statistical analysis of yield series, employing the Quade and Friedman test, suggested that the yields simulated on different soils were significantly different at $\alpha = 0.01$. All except Dencross clay (75-DCS) and the Ryerson loam (1-RYS) yield series passed the normality test. Since many decisions of land use planners and land use managers are based on the assumption that yield is normally distributed, it was considered advantageous to describe wheat yields by a family of normal probability functions. The two heterogeneous yield populations (75-DCS and 1-RYS) were dissected into components according

to a procedure described by Hald (1952), and their distributions were approximated by two normal distribution functions with parameters (μ_1, σ_1) and (μ_2, σ_2) combined in the ratio λ_1/ $\lambda_1(\lambda_1+\lambda_1=1)$.

Table 3. Parameters of wheat yield distribution functions for repesentative polygons/soils (kg/ha).

Polygon	Parameters							
	Normal distribution		Heterogeneous distribution					
	μ	σ	λ_1	μ_1	σ_2	λ_2	μ_2	σ_2
89 - RIV	2341.00	311.00						
90 -OBO*	1972.00	431.00						
75 - DCS			0.50	2296.00	214.00	0.50	1531.00	405.00
73 - RLD	2206.00	377.00						
125 - DPH*	2189.00	188.00						
53 - NDL*	2692.00	345.00						
1 - RYS			0.65	2368.00	217.00	0.35	1228.00	393.00

Note: * Killing frost occured before the wheat crop reached the maturity stage in one year out of 20 years.

Table 3 presents the parameters of the density functions as calculated from the wheat yield series. It should be noted that, in some polygons and in some years, killing frost (T $\leq$- 2°C) occurred before the wheat crop reached maturity. Such frost events were not considered in the calculation of the parameters of yield probability density functions. However, these events were included in the risk analysis.

Analyses show that wheat yield distributions differ from each other in terms of at least one of the parameters of the probability density function, either in location (μ), scale (σ), or shape (λ). The probability density curves differ not only between the two textrual classes but also within textural classes (Fig. 2a and 2b).

The Red River clay (89-RIV), an imperfectly drained soil with an artificial surface drainage network, was the most productive among the clay soils considered in the analysis. Dauphin clay (125-DPH), a soil very similar to Red River but approximately 350 km northeast, ranked second in productivity. The main difference between these two soils is amount and reliability of available solar energy to the crop. Soils in the Dauphin area warm up more slowly in the spring, delaying the seeding date and depressing crop growth. In addition, a killing frost occurred in 125-DPH in 1972 at a time when wheat development was half-way between soft dough and maturity.

On the other hand, Osborne and Dencross clay soils, which occur in the same geographic area as the Red River soil, have quite different wheat yield distributions. The yields on these two soils are generally depressed and highly variable.

On Osborne clay (90-OBO), a poorly drained soil, low yields are induced by a combination of low topographical position in the landscape, low infiltration rate (approximately 1 cm/day), presence of a water table at shallow depth, and lack of adequate surface drainage. Seeding frequently is delayed on Osborne soils because of their poor drainage and cooler spring tempera-

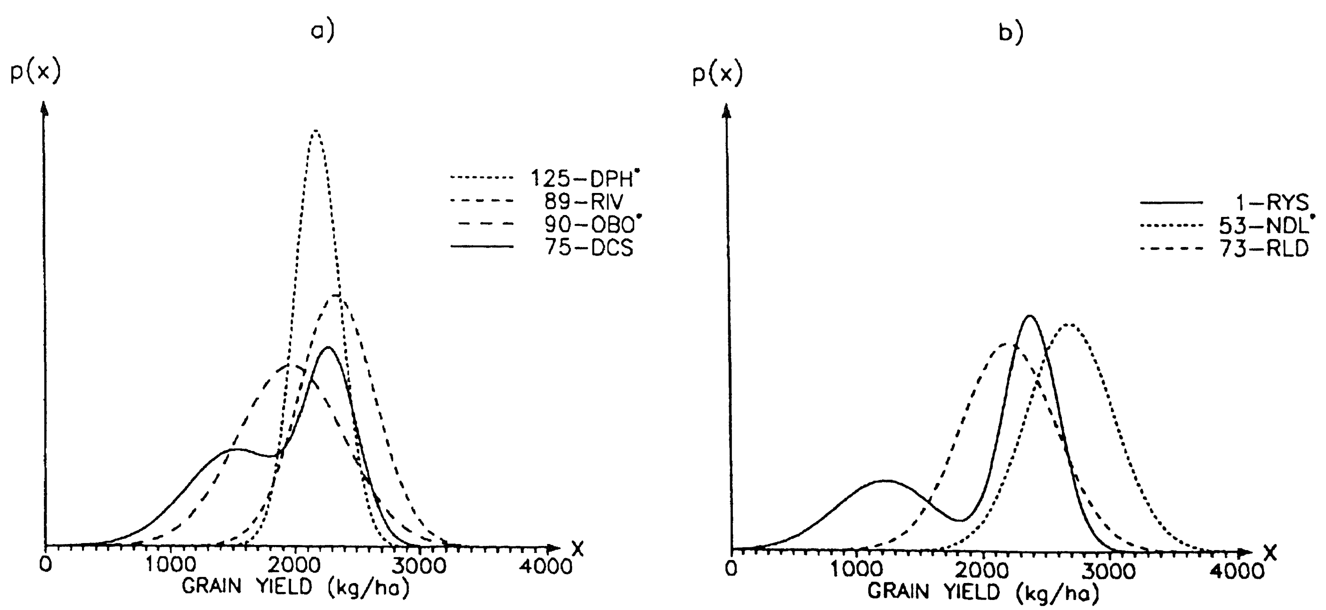


Figure 2. Probability density curves of wheat yield: a) on clay soils, and b) on loam soils.

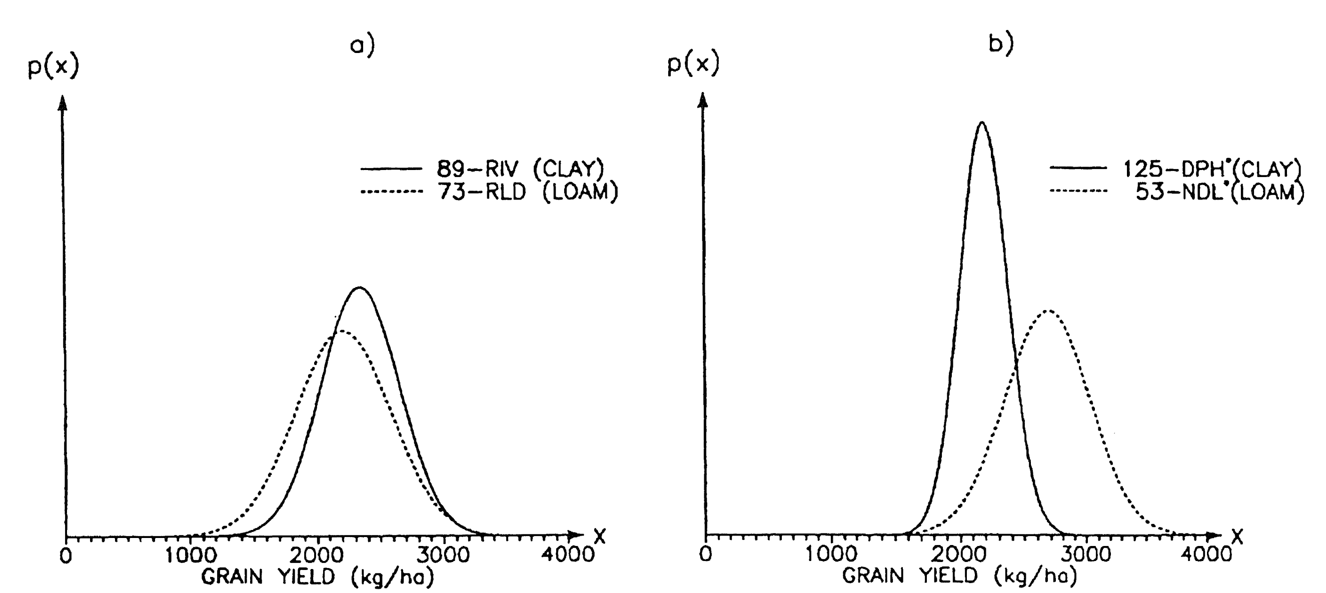


Figure 3. Probability density curves of wheat yield on adjacent clay and loam soils: a)South-Central and b)North of Black zone.

tures. Often, portions of cropped fields are flooded and wheat yields are partially lost. The likelihood of excess of water, or more precisely the lack of oxygen in the soil and consequently the depression of crop yield, is very high on Osborne soils. Also, a killing frost (T = - 4°C) occurred before the wheat crop reached maturity in 1974.

The Dencross clay (75-DCS), an imperfectly drained soil, presents somewhat better characteristics than 90-OBO. The topographical position in the landscape is higher and the infiltration rate >2 cm/day. The yield on Dencross soils, however, is depressed about 50% of the time, due to high precipitation during the growing season. In the remainder of the time the wheat yields can be as high as those obtained on the Red River clay soils. Overall, however, yield variability on Dencross clay is considerably higher than on Red River clay.

The loam soils are very similar in terms of their characteristics, but variation in yield can still be considerable (Fig. 2b). If the risk of frost is disregarded, then Newdale soil (53-NDL) outperforms the Reinland soils (73-RLD). However, in the 53-NDL polygon, a killing frost occurred before the crop reached maturity in 1982.

The yield distribution on the Ryerson loam (1-RYS) exhibits heterogeneity similar to that observed on the Dencross clay (75-DCS) but for entirely different reasons. The 1-RYS polygon is located in the driest area of Manitoba. High soil moisture deficits during the growing season on these soils resulted in depressed wheat production in 7 years out of 20 or about 35% of the time. Nevertheless, during years of high precipitation, yields on Ryerson soils can be almost as high as those on Newdale soils.

Figures 3a and 3b illustrate an overall comparison of crop production on clay soils versus loam soils for selected polygons. In south central Manitoba, the Red River clay outperformed the Reinland loam soil (Fig. 3a), whereas, near the northern edge of the Black Chernozems, the Newdale loam soil outperformed the Dauphin clay (Fig. 3b). Of particular significance was that the yield variability on both these clay soils was less than that on the loam soils included in the comparison.

Information on the long-term mean yield levels and on the shape of yield distribution curves can provide important input for risk analysis. The yield density functions can be used to quantitatively asses the probability of obtaining a given threshold yield, and this can be correlated

Table 4. The rank of land units (Polygons) based on risk of obtaining a wheat yield equal to or lower than 1783

Rank	Polygon	$P(X \leq 1783)$
1	89 - RIV	0.037
2	53 - NDL*	0.054**
3	125 - DPH*	0.064**
4	73 - RLD	0.131
5	1 - RYS	0.322
6	75 - DCS	0.366
7	90 - OBO*	0.380**

Note:
*Killing frost ($T \leq 2$°C) occurred before the wheat crop reached maturity in one out of 20 years
** Probability was calculated assuming that the yield was lost (X=0) in the years when the killing frost occurred.

with risk. Comparative production risk, which considered all clay and loam soils included in this study, was performed based on the economic break-even yield of 1783 kg/ha, calculated for Manitoba in 1987 (break-even yield is the yield necessary to cover input costs only). In this analysis it was assumed that the yield was zero in the years when a killing frost occurred before wheat crop reached maturity. Table 4 presents the probability of obtaining a yield equal to or lower than 1783 kg/ha on each soil/polygon combination.

The lowest risk of obtaining such an undesirable yield was on the Red River clay (98-RIV). On these soils one can expect yields equal to or lower than 1783 kg/ha in only about four out of 100 years; P(X=Grain yield $\leq$1783 kg/ha) = 0.037. The highest risk of obtaining this low level yield was on the Osborne clay (90-OBO), a catenary member with Red River clay. Within the agroecosystems that operate on Osborne clay, one could expect yields to be equal to or lower than 1783 kg/ha in 38 out of 100 years. The risk of obtaining this threshold yield was higher on the Reinland loam (73-RLD) than on the Red River clay but lower than on the Osborne or Dencross clay. Only a minor difference in probability was detected between the Newdale loam (53-NDL) and the Dauphin clay (125-DPH), located at the northern fringe of the Black zone. It is important to note that individual rankings are not absolute values, but they are relative to the soils included in the comparison and to the threshold yield selected as the criteria for comparison. For example, if 1168 kg/ha is taken as the threshold yield (the break-even yield for wheat calculated for Manitoba in 1985), then there are no differences in terms of risk between the Red River and Dauphin clay soils and the Newdale loam soil.

Conclusions

In general, farm type and cropping systems in the interior plains of western Canada changed with gradients in temperature and precipitation but not with soil textural classes within a region. A large proportion of available arable land in the Brown and Dark Brown zones was used for summerfallow, whereas in the Black zone continuous cropping is much more common. Spring wheat was the dominant crop cultivated in all soil zones. The largest values of total capital investment and operating expenses per unit area were found on clay soils in the Black zone, but these farms also had the highest total sales per hectare. If the ratio of sales/expenses is taken as an overall indicator of efficiency, then farms in the Brown zone were the most efficient, with those on clay soils being slightly more efficient than those on loam soils.

The wheat yield analysis showed that crop productivity varied widely not only between textural groups but also within the same soil textural class. The risk of obtaining a given threshold yield varied within catenary members of soil. Some clay soils outperformed adjacent loam soils (Red River vs. Reinland), whereas others ranked behind the loam soils (Osborne vs. Reinland, Dencross vs. Reinland, Dauphin vs. Newdale).

Defining the variability of yield based on yield frequency distributions provides a new dimension for crop yield interpretation. This information provides a more complete and realistic base for evaluating relative land quality for crop productivity. Such information can be used to assist with planning decisions and to facilitate the development of fair and equitable agricultural support programs. Results of this study indicate that only a certain number of fundamental soil characteristics, agronomic land attributes, and weather elements strongly affect productivity in the interior plains of western Canada. Considerable progress in reserch can be made by centering efforts on these key, controlling factors.

Literature Cited

Clayton, J. S., A.W. Ehrlich, B.D. Cann, H.J. Day and B.I. Marshal. 1977. Soils of Canada. Research Branch, Canada Department of Agriculture, Ottawa, Ontario. 482 p.

Conover, W.J. 1980. Practical Nonparametric Statistics. John Wiley and Sons, New York. 493 p.

Dumanski, J. and R.B. Stewart 1981. Crop production potentials for land evaluation in Canada. LRRI, Agr. Canada. Ottawa, Ontario. 80 p.

FAO, 1978. Report on the agro-ecological zones project. Vol. 1. Methodology and results for Africa. World Soil Resour. Rep. 48 Rome. 158 p.

FAO, 1984. Guidelines: land evaluation for rainfed agriculture. Soil Bull. 52. Rome. 237 p.

Hald, A. 1952. Statistical Theory with Engineering Applications. Applied Statistics, John Wiley and Sons, Inc., New York. 783 p.

Huffman, E. 1988. A description of physical and economic strategies of farming in the major soil zones of the Canadian Prairies. In Crop production risks in the Canadian prairie region in relation to climate and land resources. Dumanski, J. and V. Kirkwood (eds.) Technical Bull. 1988-5E, LRRC, Research Branch, Agriculture Canada, Ottawa, Ontario. pp 17-30.

Kirkwood, V., J. Dumanski, A. Bootsma, R.B. Stewart and R. Muma. 1989. The land potential data base for Canada. User's handbook. Res. Br., Agric. Canada, Ottawa. 53 p.

Onofrei, C. 1986. A method of Land Evaluation Using Crop Simulation Techniques. Unpubl. Ph.D. Thesis, University of Manitoba, Winnipeg. 314 p.

Onofrei, C. 1987. Land Evaluation based on wheat yield probability density functions. Technical Report, Department of Soil Science, University of Manitoba and LRRC, Research Branch, Agriculture Canada, Ottawa, Ontario. 197 p.

Williams, G.D.W. and H.N. Hayhoe. 1982. Procedures for computing and mapping Thiessen weighting factors for digitized district boundaries and climatological station latitude and longitude. Technical Bull. 1982-3E, LRRI, Research Branch, Agriculture Canada, Ottawa, Ontario. 22 p.

Reclamation Management and Techniques for Cold Entisols in Southwestern Wyoming

F.E. Parady[1] and Norman E. Hargis[2]

Abstract

Bridger Coal Company operates a 5.4 tonne year^{-1} (6.0 million tpy) surface coal mine 56 km (thirty-five miles) northeast of Rock Springs, Wyoming. Approximately 8,100 ha (20,000 acres) are under permit, with disturbance over the life of the mine projected to reach 4,050 ha (10,000 acres). Located on the western rim of the continental divide, the mine receives less than 23 cm (9 inches) of precipitation annually. Soils in the area are coarse-textured, and problems associated with elevated salinity and sodicity are encountered.

A variety of common reclamation techniques have been modified to reflect these conditions. Soil horizons are segregated during salvage operations (the surface six inches as topsoil and the balance as subsoil). Unsuitable materials are not salvaged. Direct application of soil is used to maximize native plant regeneration and conserve soil fertility. Interseeding of seeding failures has proven to be significantly more successful then chisel plowing and reseeding. Broadcast seeding has been ineffective because of strong winds, and a no till drill has been modified to handle diverse seed mixes and rock conditions. The utility of fertilization under typically xeric moisture regimes is being evaluated.

Introduction

Bridger Coal Company mines from 3.9 to 6.5 million tonne year^{-1} (4.4 to 6.5 million tpy) of subbituminous steam coal at its surface strip mine located 56 km (thirty-five miles) northeast of Rock Springs in southwestern Wyoming. The mine is adjacent to the continental divide at elevations ranging from 2,073 to 2,150 m (6,800 to over 7,100 feet.)

Mean annual precipitation is 0.22 m (8.8 inches), and the average number of frost free days is 100 (Bridger Coal Company, 1980). Bridger's permit to mine encompasses nearly 8,100 ha (20,000 acres), with 2,000 ha (4,950 acres) disturbed and over 520 ha (1,280 acres) or 26% reclaimed to date (Bridger Coal Company, 1988). Life of mine disturbance is projected to reach approximately 4,050 ha (10,000 acres).

Reclamation feasibility on arid lands has been questioned since the resurgence of the western coal mining industry in the early 1970s. The National Academy of Sciences (1974) suggested that 0.25 m (ten inches) of precipitation was necessary to sustain revegetation. Bridger Coal has developed or modified a variety of reclamation equipment and techniques to reflect local conditions and provide the foundation for successful reclamation. The following information is based upon observations during the past 8 growing seasons.

[1]Manager, Safety and Reclamation, Bridger Coal Company, P.O. Box 2068, Rock Springs, Wyoming, 82901.

[2]Reclamation Coordinator, Bridger Coal Company, P.O. Box 2068, Rock Springs, Wyoming, 82901.

Soil Environment

An Order One soil survey was conducted in 1978 which identified 25 series. Soils are predominantly Entisols (Table 1). Torrifluvents occur in alluvium along major ephemeral drainages. Torripsamments occur on sand dunes. Torriorthents occur in smaller ephemeral drainages and on slopes and uplands. The typical thickness of A horizons is 0.05-0.13 m (2-5 inches). Normally C horizons are immediately below the A horizon; AC horizons are rare.

Soil formation has been limited and is closely related to local geology. Soils are derived from Cretaceous calcareous shale, Tertiary sandstone and shale, aeolian sand, and alluvium (Bridger Coal Company, 1980).

The temperature regime is frigid. The 25 soil series support 7 range sites: saline upland, shallow loamy, loamy, sands, saline lowland, impervious clay, and saline subirrigated. Principal plant communities are gardner saltbush, big sagebrush-wheatgrass, greasewood, and big sagebrush-rubber rabbitbrush. Vegetative cover is less than 25%. Rock outcrops with little or no soil cover are common. Slickspots, which are saline-sodic and unsuitable as soil, occur in some major drainages.

Table 1. Major taxonomic groupings of soil series in Bridger Coal Company mining permit, Sweetwater County, Wyoming.

Order	Great Group	Series
Aridisols	Natrargids	Westvaco
	Calciorthids	Cambarge
		Pepal
	Camborthids	Sage Creek
Entisols	Torrifluvents	Quealman
		Laney
		Chrisman
	Torriorthents[1]	Haterton
		Horsley
		Huguston
		Leckman
		Rock Springs
		Terada
		Thayer
		Garsid
		Monte
		Dinco
		Dines
		Boltus
		Tasselman
		Winton
	Torripsamments	Corlett
		Fandaly

[1]Includes one unnamed series.

Specifications for soil management are based on soil chemical and physical properties. Soils are salvaged for use in reclamation to the paralithic contact or to the depth of elevated salt, sodium, or other unsuitable property. This insures a suitable soil environment for seedlings and growing plants.

Soil Management

Soils on the mine site are typically Entisols, coarse textured, with an average pH of 7.5 to 8.0 and electrical conductivity in the 4.0 to 6.0 ds m^{-1} range. Problems associated with elevated salinity, sodicity, and boron levels are encountered. Bridger Coal has implemented a soil management program to assure proper use of soil resources.

First, a staking program is used on the highwall to identify unsuitable native soils. Unsuitable materials are not be moved onto recontoured spoil unless soil heterogenity is specifically desired. Suitability is defined in accordance with state guidelines (Table 2).

During soil stripping operations, soil horizons are segregated, the surface 15 cm (six inches) as topsoil and the balance as subsoil. Soils range from 15-60 cm thick (six inches to sixty inches), with a mine wide average of 38 cm (fifteen inches). Revised permit language has been approved by the state to allow for variable soil application thicknesses from 30 cm (twelve inches) on the ridge positions to 76 cm (thirty inches) in lowland positions. Variable soil depths provide a foundation for vegetative diversity.

Stripped soil is either stockpiled or hauled directly onto a completed regraded area. Stockpiles are contoured to allow farming operations on the side slopes and to minimize erosion.

Direct application of soil is a key element in achieving diversity elements of bond release criteria. Soil is picked up with scrapers and transported across or through the pit, and then placed directly on ripped recontoured spoil. Bridger Coal initiated direct application of soil in 1976, and has succeeded in using this preferable method on over 240 reclaimed hectares (600 acres) to date. The increased cost of longer hauls associated with direct application is offset by elimination of double handling incurred by stockpiling. This technique maximizes native plant regeneration and conserves soil fertility. Bridger Coal has also completed soil application to 70 ha (175 acres) using stockpiled material as a subsoil covered with a 0.3 m (six inch) surface application hauled directly form the highwall.

Specifically, ten species have volunteered from direct applied soil. These species include big sagebrush (*Artemisia tridentata*), Sandberg's bluegrass (*Poa sandbergii*), Fendler's bluegrass (*Poa fendleriana*), greasewood (*Sarcobatus vermiculatus*), rubber rabbitbrush (*Chrysothamnus nauseosus*), Plains wallflower (*Erysimum asperum*), scarlet globemallow (*Sphaeralcea coccinea*), and scarlet gilia (*Ipomopsis aggregata*). Direct application of soil aids in returning the shrub and forb components of the plant community, as well as in establishing understory species of the grass component.

Overburden/Parting Management

The objective of the overburden characterization and handling plan is two fold. First, to insure that the surface 1.2 m (four feet) of mate-

Table 2. Selected Criteria for Assessing Suitability of Materials for Use Within the Root Zone.

Parameter	Suitable	Unsuitable
Acid-Base Potential (A/B) (Tons $CaCO_3$ per 1,000 tons)	> -5.0	< -5.0
Boron (ppm)	< 5.0	> 5.0
Electrical Conductivity (ds m^{-1})	< 8.0	> 8.0
Exchangeable Sodium percentage (%)	< 20.0	> 20.0
Molybdenum (ppm)	< 1.0	> 1.0
pH	5.0 - 9.0	< 5.0, > 9.0
Saturation percentage (%)	25 -80%	< 25, >80%
Sodium Adsorption Ratio	< 10.0	> 10.0
Texture	< 50% clay, < 85% sand,	> 50% clay > 85% sand
Total Organic carbon (%)	< 10.0	> 10.0

rial is suitable for plant growth. Second, to insure that unsuitable materials are not placed in surface drainages, where it could be eroded. To meet these objectives, we employ the following techniques:

1. Drilling on a 65 ha (160 acre) spacing. Laboratory analysis has been completed on the entire area. This program defined the specific problems (salinity, sodicity, and boron) and zones of unsuitable materials.
2. Unsuitable materials comprising less than 15% of the total volume are mixed during mining with draglines.
3. Supplemental stripping equipment is used to extract unsuitable materials, primarily interburden, and place them on the pit floor, below the regraded surface but above the anticipated potentiometric surface.
4. The dragline swing may be modified to place material lower within the spoil peak for subsequent burial.
5. Following recontouring, regraded spoils are sampled on 120 m (400 foot) grid in 0.6 m (two foot) intervals to a 1.2 m (four foot) depth. Samples are analyzed for pH, EC, ESP or SAR, molybdenum, boron, and acid/base potential. Results are evaluated using the criteria in Table 1.
6. If unsuitable materials are identified, they are either covered or relocated.

Since the fall of 1981, samples have been taken and analyzed to provide a basis for evaluating the suitability of root zone materials prior to soil application. Results have been reported in detail in the 1982 through 1988 Annual Reports. To date, 24 areas comprising nearly 220 ha (540 acres) have been evaluated. A total of 516 samples have been taken. Less than seven percent of these samples have had unsuitable value for any parameter, and less than 0.8 ha (two acres) have been withheld from soil application on the basis of spoil unsuitability. These results clearly demonstrate the guaranteed cover plan has been successful in accomplishing proper materials placement.

Farming Operations

The goal of reclamation is set by statute under the Surface Mining and Reclamation Act of 1977 to establish a diverse, native plant community capable of regenerating itself. Diverse techniques must be employed to achieve diversity of species in a reclaimed plant community. Seed drills used initially in reclamation, specifically the Laird Rangeland drill, were not capable of handling fluffy, trashy native seeds such as winterfat (*Ceratoides lanata*). Consequently, only 5 or 6 species were seeded in early reclamation efforts, generally wheatgrasses and fourwing saltbush.

To remedy this, Bridger Coal Company purchased a Tye "Pasture Pleaser" no till seed drill. The drill has three seed boxes, with fairly standard wheatgrass and legume boxes. The third, a shrub box, is specifically equipped with agitator discs and larger picker wheels to handle trashy seed. In addition, the large seed tube from this box can distribute seed across the entire furrow. The result is variable planting depths, including shallow planting depths that are desirable for most of these native species. Bridger Coal has therefore been able to use 18 to 20 species in each of its seed mixes, corresponding to four range sites: shallow loamy, sands, saline upland, and saline lowland. Big sagebrush (*Artemesia tridentata*) was successfully established with this technique in several fall seedings. Studies by DePuit and Coenenberg (1979) have indicated that increasing the number of species in a seed mix increases the diversity of the resulting plant community.

An additional technique that has proven successful is interseeding. Interseeding involves seeding with a no till directly into an existing reclaimed surface, rather than chisel plowing and reseeding. The advantage lies in minimizing disturbance to the soil and in keeping existing vegetation intact. During the fall of 1981, portions of a reclaimed area were either interseeded or chisel plowed and reseeded. By 1984, the interseeded area showed 180 desirable plants m^{-2}, compared with 53 desirable plants m^{-2} on the area chisel plowed and then reseeded (Bridger Coal Company, 1984). Interseeding can also be useful as it provides a second age group of plants within the community.

Broadcast seeding has had limited success at Bridger Coal, primarily because of wind. Broadcast seeding is intended to provide the shallow planting depth necessary for native species, as well as improving reclamation aesthetically by eliminating the appearance of drill rows. A modified broadcast seeder was used in 1981 on 150 acres. Average first year seedling density resulting from the broadcast seeder was 15.2 seedlings m^{-2}. Average first year seedling density on 53 ha (130 acres) seeded with a drill in

1981 was 47.5 seedlings m^{-2} (Bridger Coal Company, 1984), a 300% difference.

Several changes in mulching operations have significantly improved productivity. First, a round bale buster was purchased to replace both a blower type mulcher and a tub mulcher. The blower was labor intensive, requiring a tractor operator, two hay handlers, and blower operator. The tractor operator can self-load the bale buster from the cab, eliminating the need for three people. Second, large (500 - 725 kg) (1100 - 1600 pound) round bales are used, eliminating the handling involved with small bales. Third, a 4.6 m (15 foot) working width flexible crimper with hydraulically operated gangs and transport wheels was put into service, replacing a small 1.8 m (six foot) crimper. A final improvement in the mulching operation was replacing straw with grass hay. Hay adheres to the surface better.

Shrub Establishment

Proposed state regulations require establishment of one shrub per square meter in a mosaic pattern on 10% of the mine's reclaimed area. This standard has been met on 20 ha (48 acres) or 6% of reclamation to date and has nearly been met on several additional areas.

Three species of sagebrush (*Artemisia tridentata subsp. wyomingensis, Artemisia cana, Artemisia frigida*), fourwing saltbush (*Atriplex canescens*), Gardner's saltbush (*Atriplex gardneri*), winterfat, rubber rabbitbrush, greasewood, and spiny hopsage (*Grayia spinosa*) are currently used in different seed mixes to promote shrub establishment for wildlife habitat. Fourwing saltbush, Gardner's saltbush, and winterfat have been especially successful in reclamation seedings.

Direct application of soil also maximizes shrub establishment by increasing the survival of propagules and vesicular arbuscular mycorrhyzae that remain in the soil at the surface of reclamation. Wyoming big sagebrush, rubber rabbitbrush, and greasewood have been established with the technique.

Irrigation

A cooperative research project with the University of Wyoming has been completed to assess the establishment of a predominantly native, diverse seed mix under irrigation. The objectives of the research included determining optimum irrigation rates for initial vegetation establishment; determining optimum seasonal scheduling and duration; and defining interactive effects of varied treatments on initial and ultimate vegetation density, productivity, species composition, and diversity (Vincent et al, 1986).

Fertilization

Analysis of a poor reclamation area first seeded in 1981 revealed total nitrogen and available phosphorous levels (.03% N and 2.1 ppm P) below desirable plant available levels (Bridger Coal Company, 1981). Fertilizer at 170 kg/ha (150 lbs/acre) of 18-46-0 was applied in the spring of 1984 and the area was then interseeded. Seedling establishment from the interseeding appears satisfactory and established plant vigor appears improved. However, the utility of fertilizer in the region is probably limited to average or above average precipitation years.

Summary

Many of the initial concerns over reclamation feasibility in a semi-arid desert environment have been laid to rest. Improvements have occurred in soil management, shrub establishment, and farming operations. Experiments are underway with various techniques such as irrigation and fertilization.

Wyoming Department of Environmental Quality personnel have ocularly evaluated all reclaimed areas at Bridger Coal annually since 1982. The percentage of reclamation rated good or fair increased from 38.2% in 1982 to over 90% in 1988. The percentage of disturbance reclaimed has doubled in less than five years, from 13.3% in 1979 to 26% in December of 1988. Reclamation has been successfully achieved in areas receiving less than 0.25 m (ten inches) of precipitation.

Literature Cited

Bridger Coal Company, 1980. Jim Bridger Mine Base Document, Wyoming Permit to Mine No.338-C. Volume 2: Mine Plan and Volume 17: Soil Survey Report. P.O. Box 2068, Rock Springs, Wyoming.

Bridger Coal Company, 1981. Jim Bridger Mine 1981 Annual Report, Permit to Mine No. 338-C. Volume 3: Irrigation. P.O. Box 2068, Rock Springs, Wyoming.

Bridger Coal Company, 1983. Jim Bridger Mine 1983 Annual Report, Permit to Mine No. 338-T1. Volume 1. P.O. Box 2068, Rock Springs, Wyoming.

Bridger Coal Company, 1984. Internal correspondence-Vegetation Monitoring. P.O. Box 2068, Rock Springs, Wyoming.

DePuit, E.J. and J.G. Conenenberg, 1979. Methods for establishment of native plant communities on topsoiled coal stripe mine spoils in the Northern great Plans. Reclamation Review 2: 75-83.

National Academy of Sciences, 1974. Rehabilitation potential of western coal lands. Ballinger Publishing Co., Cambridge, Massachusetts.

Vincent, R.V., E.J. DePuit, J.L. Smith, J.A. White, and F.E. Parady. 1986. Design of temporary irrigation for plant establishment on arid coal mined lands in Wyoming. Pages 45-47 in 39th Annual Report, Vegetative Rehabilitation and Equipment Workshop, U.S. Department of Agriculture, forest Service, Misoula, Montana.

Wyoming Department of Environmental Quality, 1984. Guideline No.1-Soils and Overburden. Cheyenne, Wyoming.

Vertisols of New Caledonia: Distribution; Morphological, Chemical, and Physical Properties; and Classification

P. Podwojewski and A.G. Beaudou[1]

Abstract

The Vertisols of New Caledonia are found on the west coast of the main island, where a dry and varied climate prevails. They occur in the lower parts of soil catenas, overlying basic rocks, or on old old alluvial terraces. The chemical composition of the parent material determines the chemical properties of the soils. Three kinds of vertisol have been identified: calci-magnesic vertisols, hypermagnesic vertisols, and sodic and acid vertisols. Some forms, however, show characteristics intermediate between the three categories. This paper describes the vertisols in terms of their morphological and analytical features and proposes a means of classification.

Introduction

New Caledonia is situated slightly north of the Tropic of Capricorn (Fig. 1). Vertisols there have been described as "tropical black clays" in the past (Tercinier, 1953). They cover an approximate total area of 100,000 ha (Latham et al., 1978). Compared to the surface area of the Territory as a whole (16,000 km^2), this area is relatively small, but vertisols represent nearly 50% of the land suitable for mechanized agriculture. These deep soils are present only on the west coast of New Caledonia (Podwojewski, 1984; Podwojewski, Beaudou, 1987) (Fig. 2).

Vertisols have a succession of three horizons (A.F.E.S., 1988). They include a surface horizon (Vs), an intermediate horizon (V) with a polyhedral or prismatic structure and a number of shiny or furrowed surfaces, and a vertic horizon (Vv) with a wedge-shaped, rhombohedral or

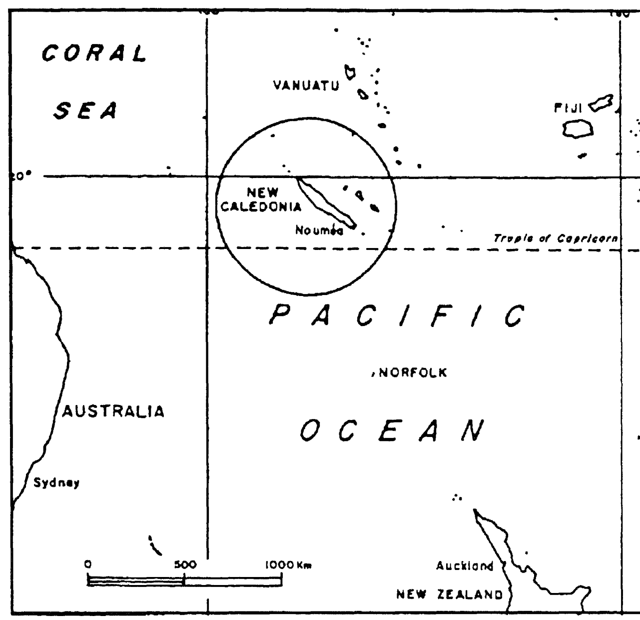
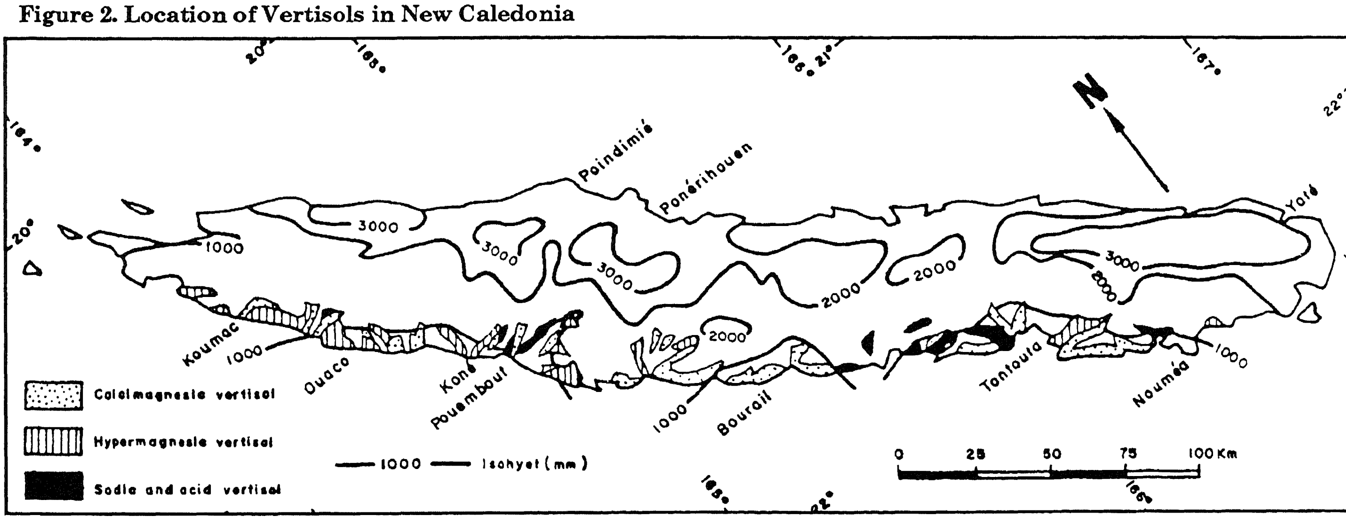


Figure 1. Location of New Caledonia

Figure 2. Location of Vertisols in New Caledonia



[1]Podwojewski, Orstom, 70-74 route d'Aulnay, 93140-Bondy (France). A.G. Beaudou, Jalan Tampak Siring, 10, Cipete, Jakarta Selatan (Indonesia).

sphenoidal structure. The high content of smectites in these horizons is responsable for the vertic structure.

Formation

Parent material, landscape position, and climatic conditions influence the formation of vertisols in New Caledonia. Most vertisols generally developed from three major groups of parent material. These are relatively basic rocks that are fairly rich in alkali/earth cations (calcium or magnesium). The weathering of these rocks produces smectite type clays.

The first of the two petrographic groups includes flysches, pelites, limestones and basalts, dolerites, and gabbros, basic rocks in which %SiO varyies from 50 to 55%. The dominant base is calcium, closely followed by magnesium. In some cases, the magnesium may , however, be slightly dominant. These rocks contain little potassium or sodium.

Serpentinites (associated with peridotites) constitute the second group. These are ultrabasic rocks containing less than 50% SiO_2. The only base present is magnesium.

The third group is composed of graywackes and schists. These rocks are more acid than the other groups and contain more than 55% SiO_2. The dominant bases are magnesium and sodium. Little calcium or potassium is in these rocks.

Location

Vertisols are on the plains of western New Caledonia at an elevation less than 100 m. (Fig. 3). They always are located in the lower portions of the landscape (Beaudou et al., 1983). They occur on the lower part of toposequences formed on materials derived directly from the subjacent rock or on colluvia. The soil sequence is as follows: an immature erosion soil, a vertic brown soil, and a vertisol. The transition to vertisol is most evident in places where the slope is gentle and where external drainage is slow.

Vertisols also occur on the old alluvial terraces of the main water courses on the west coast of New Caledonia. These are commonly the highest terraces. External drainage is very limited or non-existent. There is a continuum between vertisols formed on the substratum (lithomorphous vertisols) and vertisols forming in old alluvia (topomorphic vertisols). Gilgaï microrelief is not observed in New Caledonia.

Climate

New Caledonia is characterized by a semi-hot oceanic tropical climate. Vertisols occur only on the plains of the west coast, leeward of the south-easterly tradewinds. In these areas, mean annual rainfall is usually less than 1,000 mm, with minimum values below 800 mm.

There is a great deal of annual and seasonal variability rainfall. The wet season, however, spans mid-December to the end of March, and the dry season runs from September to November.

The mean annual temperature is approximately 23°C. In the hot season (February), it rises to 26°C. In August, during the cool season, the mean temperature drops to 19°C.

The soil moisture balance is negative. Only in July and August (the southern winter) is there a positive moisture balance. From September to November, the balance is distinctly negative.

Types of Vertisols in New Caledonia

Three groups of vertisols may be distinguished by their morphological and chemical properties. The calcimagnesic vertisols are formed from limestone, flysch, or basalt. Their extent is more than 55,000 ha. The hypermagnesic vertisols are formed from serpentinites or serpentinite colluvia associated with colluvia derived from peridotites. They cover about 35,000 ha. The sodic and acid vertisols are formed in siliceous pelites or graywackes. Their extent is less than 10,000 ha.

Morphological Features

Tables 1 and 2 summarize the morphological features of these soils.

<u>Calcimagnesic Vertisols</u>

Salts accumulation generally occurs at depth of approximately 100 cm, within the vertic horizon (Vv). Commonly a substantial accumulation of gypsum crystals (Vgy) is observed. In places, gypsum accounts for more than 25% of the dry weight of the horizon. These crystals are lenticular, commonly between 0.5 and 2 cm in size but sometimes larger (Podwojewski, 1984).

In the vertisols located on flat areas, the Vgy horizon may be underlain by substantial dark gray or black accumulations of manganese (VMn). The thickness of this layer is commonly

Figure 3. Tamoa Sequence, near La Tontouta (cf. Fig. 2). Basaltic hill: calimagnesic vertisol on the lower part of the slope. High alluvial terrace: hypermagnesic vertisol derived from serpentinite and peridotite.

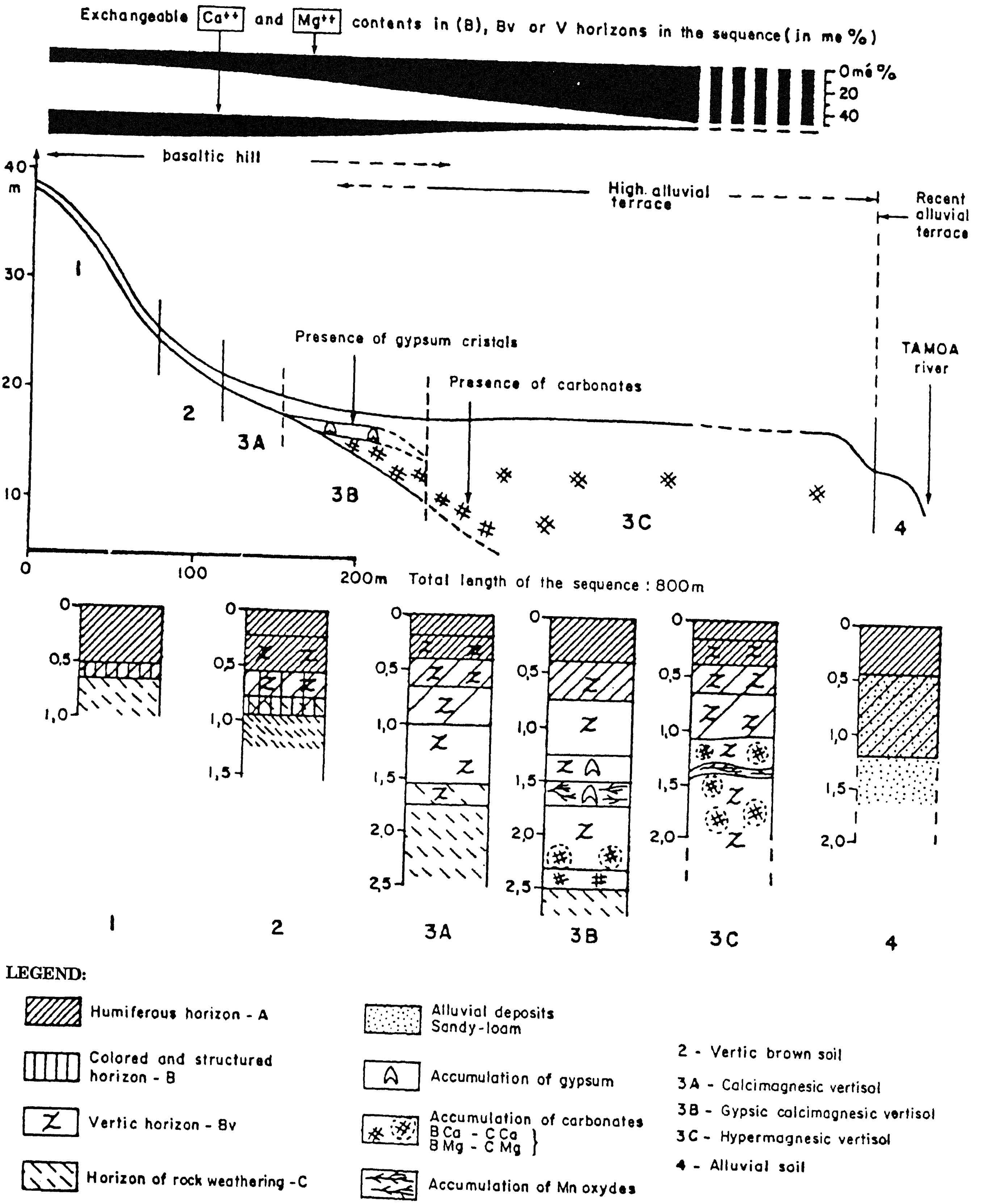


Table 1 - Morphological Features in the Upper Part of the Profile of the Calcimagnesic (A), Hypermagnesic (B), and Sodic and Acid (C) Vertisol Types

HORIZON	TYPE	DEPTH (cm)	COLORS	TEXTURE	STRUCTURE	PEDOLOGICAL ACCUMULATIONS
A_1 Surface horizon	A	0 - 30	10YR 3/1 very dark gray	Clay	fine to medium angular blocky	few manganese concretions diameter < 0.5 cm
	B	0 - 30	10YR 3/1 to 10 YR 2/0 black	Clay	fine to medium angular blocky	few manganese concretions diameter < 0.5 cm
Surface degraded horizon	C	0 - 10	10YR 4/1 dark gray	Silt loam	fine angular blocky	many manganese concretions diameter 0.1 to 1.0 cm sometimes 10% of the volume
A_{12} Surface degraded horizon	C only	10 - 25	10 YR 3/1 very dark gray	Clay loam	angular blocky medium	few manganese concretions diameter 0.1 to 1.0 cm
AB Intergraded horizon	A	25 - 50	often mixed colors (infilled cracks) 10YR 3/2 very dark grayish brown	Clay	coarse angular blocky to coarse prismatic	sometimes nodules of $CaCO_3$ < 2 cm rare manganese concretions < 0.5 cm
	B	30 - 50	2,5Y 3/2 very dark grayish brown	Clay	few pressure faces, few slickensides < 10 cm long	rare manganese concretions < 0.5 cm
	C	30 - 50	10YR 3/2 to 4/2 dark grayish brown	Clay		rare manganese concretions < 0.5 cm
BV Vertic horizon low organic matter: infilled cracks or infilled worm channels	A	50 - 70	10YR 3/3 to 4/4 dark brown	Heavy Clay	parallelepiped structure, intersecting slickensides: 20-40° from horizontal > 15 cm long	rare manganese concretions < 0.5 cm
	B	50 - 70	2,5Y 4/2 to 4/4 dark grayish brown			rare manganese concretions < 0.5 cm
	C	50 - 70	7,5YR 3/4 to 4/4 dark brown			rare manganese concretions < 0.5 cm

between 10 and 15 cm. The manganese oxides occur in the form of dendrites and cutans, and also in the form of nodules of varying degrees of cementation. The manganese oxide contents do not exceed 5% of the weight of the dry soil. It is likely that the presence of these elements is due to the existence of a former water table. The gypsum crystals in and under this VMn horizon are agglomerated and form "sand rose" type formations. The gypsum crystals then disappear fairly rapidly with depth.

Calcium carbonates occur below the gypsum. They gradually appear in the form of "powdery volumes," very small pockets, nodules, and crusts of varying degrees of cementation. Carbonate contents increase with depth.

Hypermagnesic Vertisols

These vertisols are similar to calcimagnesic vertisols with respect to:

- the occurence of carbonates below the VMn horizons in varying forms, from very small pockets to crusts, depending on depth
- the presence, at depth, of a horizon where manganese oxides have accumulated.

They differ in these ways:

- total absence of gypsum crystals
- presence of accumulations of magnesium carbonates (magnesite or giobertite), which are rarely present in continuous form (crusts).

Other differences may also be noted but these are often much less marked. Hypermagnesic vertisols appear blacker at the surface (10YR2/1 to 2.5Y2/0) than the calcimagnesic vertisols and the sodic and acid vertisols. The vertic horizons of the hypermagnesic vertisols are more olive brown in color (2.5Y4/4 to 5/6) than the calcimagnesic vertisols (10YR4/4 to 5/6).

Sodic and Acid Vertisols

Their essential features occur in the surface horizon, which is grey in color (10YR4/1 or 5/1) and in texture is much less clayey (silt loam or loam). It also often includes abundant concretions of manganese oxide ("pellets").

The vertic horizons possess a less distinct wedge-shaped structure. The cohesion is very strong. The coloring is usually brighter (7.5YR to 5YR4/4 to 5/6).

The sodic and acid vertisols lack sulfate concentrations. Carbonate accumulations are few. They are calcic and magnesic, occurring in the form of highly friable nodules and pockets.

The accumulation of manganese oxides at depth is less marked. These oxides occur in the form of nodules, dendrites, and very small pockets.

Table 2 - Morphological Features in the Lower Part of the Profile of the Calcimagnesic (A), Hypermagnesic (B), and Sodic and Acid (C) Vertisol Types

HORIZON	TYPE DEPTH (cm)	COLORS	TEXTURE	STRUCTURE	PEDOLOGICAL ACCUMULATIONS	CONSISTENCE
V v or (B)	A 90 -150+	10 YR 4/4, 5/4, 5/6 yellowish brown	Heavy Clay	parallelepiped structure, many intersecting slickensides: 20-40° from horizontal > 15 cm long	rare manganese concretions < 0.5 cm rare soft carbonate	plastic (wet) sticky firm (moist)
V Gy or (B)Gy NOT ALWAYS PRESENT	A 90 -150+ only				lenticular gypsum crystals 0.5 to 2.0 cm long, rare 5.0 cm sometimes > 20% weigh of air dried soil In VMn horizon and below: agglomerate crystals into spheric accumulations 2 to 15 cm diameter	
V Ca or (B)Ca Under V Gy When V Gy is present	A 90 -150+ frequent C rare				In the upper part of the horizon: soft carbonate in spheroid volumes 2-10 cm diameter. In the lower part: little soft concretionary carbonate; many nodules 1-3 cm diameter.	
V v or (B)	B 90 -150+	2,5 Y 4/4, 5/4, 5/6 light olive brown	Heavy Clay	parallelepiped structure, many intersecting slickensides: 0-40° from horizontal > 15 cm long	are manganese concretions < 0.5 cm rare soft Mg carbonate	plastic(wet) very sticky firm (moist)
V Mg or (B)Mg	B 90 -150+ frequent				In the upper part of the horizon: soft Mg carbonate in spheroid volumes 2-10 cm diameter. In the lower part: little soft concretionary carbonate; many white nodules 1-3 cm diameter of magnesite	
V v or (B)	C 90 -150+	7,5 YR 4/4, 4/6 5 YR 4/4, 4/6 strong brown to yellowish red	Heavy Clay	parallelepiped structure, < 0.5 cm intersecting slickensides: 20-40° from horizontal > 15 cm long	rare manganese concretions rare soft Ca carbonatee	slightly plastic (wet) slightly sticky very firm (moist)
V Mn only in old alluvial terraces	A or B depth varies, 10 to 15 cm thick	black	Heavy Clay	same structure as in Vv for type A or B soils	manganese concretions and manganans on the slickensides and around the pores (on tubular and planar voids)	

Analytical Features

Table 3 presents the chemical analysis of these soils.

Mineralogy

The calcimagnesic vertisols consist of well crystallized montmorillonite (with a little quartz).The hypermagnesic vertisols feature the presence of ferriferous smectite of the bowlingite variety (Latham & al, 1978), lacking aluminium. The degree of crystallinity appears to be higher (clearer peaks) than in other vertisols.

The sodic and acid vertisols consist of montmorillonite associated with interstratified illite and montmorillonite.

Texture

All the vertic horizons have a clay content over 50%. However, the presence of carbonates and manganese oxides brings about a relative decrease in the clay content of these horizons.

The surface horizons are generally less clayey. In the sodic and acid vertisols, the impoverishment in the clay content (or leaching) at the surface is very evident. However, this clay does not redistribute itself at depth in the form of argillans in an accumulation horizon (Denis, Mercky, 1979).

pH

The surface horizons of all vertisols in the study area have a pH between 5.5 and 6.0. At depth, the values vary, depending on the type of vertisol.

Calcimagnesic vertisols with high salt accumulations have a pH which varies with the nature of these accumulations: 5.5 where gypsum is present, 7 to 8 where carbonates are present.

The hypermagnesic vertisols, lacking sulfates, but rich in magnesium carbonates at depth, feature regularly increasing pH values which may reach or exceed 8 in the giobertite (magnesite) horizons.

The sodic and acid vertisols have a pH lower than 6.0 and in some places even less than 5.0. Under such conditions, exchangeable aluminium may be present.

Exchangeable Bases and Exchange Capacity

The method for extracting exchangeable bases recommended by Tucker (1985 b.) using ammonium chloride (at pH 7 in Noumea) has

Table 3 - Chemical Analysis of the Calcimagnesic (A), Hypermagnesic (B), and Sodic and Acid (C) Vertisol Types

TYPE OF VERTISOL	A	B	C	C	A	B	C	A	B	C	C
HORIZON	A_1	A_1	A_{11}	A_{12}	AB	AB	AB	BGy	BMg	B	BCa
DEPTH (cm)	0-30	0-30	0-12	12-35	40-60	40-60	40-60	100-140	80-120	80-100	195-240
GRANULOMETRY %											
Clay	46.5	44.6	24.0	37.6	47.6	59.2	53.9	70.9	54.4	49.3	64.4
Fine silt (0.002-0.02mm)	21.8	17.6	30.0	20.5	19.5	10.6	18.5	12.1	11.4	17.2	11.8
Coarse silt (0.02-0.05mm)	11.0	9.2	22.5	11.2	9.7	6.1	9.9	7.6	4.0	10.0	5.7
Fine sand (0.05-0.2mm)	10.4	16.7	16.5	17.3	11.4	11.8	16.3	4.9	14.7	21.8	5.9
Coarse sand (0.2-2mm)	5.6	7.4	3.5	12.7	11.0	9.3	2.0	5.8	13.8	2.0	10.5
MOISTURE (g/100g) at pF											
2,5	41.0	47.4	33.7	31.7	42.7	62.5	39.2	46.2	54.3	39.2	44.0
4,2	23.9	29.8	12.2	16.4	23.3	40.3	22.3	26.3	35.9	18.5	28.2
ORGANIC MATTER g/100g #											
C	3.04	2.13	2.337	0.847	0.84	0.85	0.489				
N	0.184	0.182	0.141	0.100	0.073	0.070	0.074				
Organic Matter	5.2	3.7	4.0	1.5	1.4	1.5	0.8				
C/N	16.5	11.7	16.6	8.5	11.5	12.1	6.6				
pH H_2O 1:2.5	5.9	6.2	5.3	5.1	6.2	7.6	4.5	5.8	8.4	4.6	8.2
pH KCl	4.8	5.0	4.7	4.2	5.1	6.2	3.6	4.8	6.8	3.7	7.1
EXCHANGE COMPLEX cmol(p^+)/kg *											
Ca^{++}	12.1	1.8	4.4	4.3	11.2	0.23	3.6	11.9	0.34	3.5	19.4
Mg^{++}	15.0	40.20	3.8	7.3	16.2	66.90	10.5	23.6	59.20	10.6	28.2
K^+	0.84	0.08	0.2	0.1	0.11	0.04	0.14	0.14	0.03	0.12	0.12
Na^+	0.65	0.48	0.34	2.9	2.8	1.30	4.7	4.69	1.40	5.1	6.16
Σ Cations	28.59	42.56	8.74	14.5	30.31	68.47	18.94	40.33	60.97	19.32	53.88
C.E.C.	36.2	49.0	17.7	21.0	34.4	61.0	23.4	41.6	57.9	22.5	46.2
Base saturation %	78.9	86.9	49.4	69.0	88.1	SAT.	80.9	96.9	SAT.	85.9	SAT.
Total phosphorus P_2O_5 mg/kg	480		430	420	130		150				
PERCHLORIC TOTAL ANALYSIS g/100g #											
Loss on ignition	10.44	10.57	6.6	6.9	7.72	9.47	6.6	7.74	7.94	5.5	8.43
Residue	38.96	54.98	59.4	47.9	41.30	56.46	45.4	28.40	61.94	49.6	29.14
SiO_2	32.68	28.86	25.0	26.6	29.60	38.26	28.7	34.00	39.80	26.9	35.42
Al_2O_3	6.31	2.87	3.6	9.1	9.07	2.65	11.7	11.15	1.97	9.8	10.58
Fe_2O_3	6.35	22.88	2.4	6.3	7.35	22.02	6.0	7.58	17.16	5.6	8.72
MnO_2	0.58	1.00	0.77	1.5	1.24	1.23	0.05	0.04	0.42	0.03	0.22
TiO_2	0.87	0.18	0.43	0.60	0.84	0.16	0.72	0.86	0.10	0.67	0.98
CaO	0.48	0.19	0.15	0.09	0.39	0.10	0.07	4.20	0.13	0.07	2.94
MgO	0.76	4.22	0.23	0.47	0.84	5.34	0.63	1.22	9.07	0.62	2.58
K_2O	0.08	0.02	0.13	0.20	0.06	0.01	0.30	0.11	0.01	0.36	0.15
Na_2O	0.08	0.04	0.10	0.16	0.16	0.06	0.29	0.32	0.07	0.34	0.45
SiO_2/Al_2O_3 mol.	8.8	17	11.9	5.0	5.5	24.5	4.2	5.2	34.2	4.7	5.7
Cr_2O_3		2.32				1.11			0.52		
Mg/Ca	1.2	22.3	0.86	1.7	1.4	91	2.9	2.0	174	3.0	1.4
Na/CEC % (ESP)	1.8	1.0	1.9	13.8	8.1	2.1	20.1	11.3	2.4	22.6	13.3

* Exchangeable bases on vertisols of type A have been extracted by B.M. Tucker's methodology: Ammonium chloride at pH 7.0
\# g/100g of dried soil at 105°C

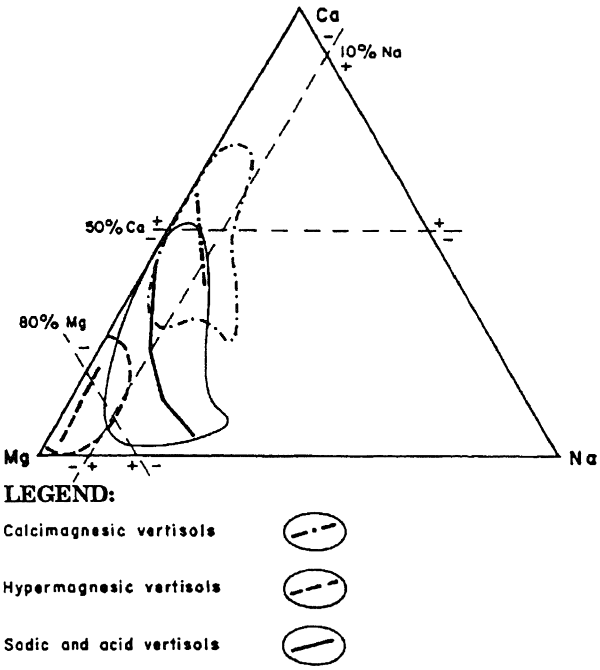
proved the most suitable for the vertisols of New Caledonia. This method restricts the influence of soluble salts in the extraction of exchangeable bases, mainly in the gypsum and carbonate horizons.

Calcimagnesic vertisols have an exchange capacity that is approximately 40 me/100g soil (see figure 4). They are base saturated in the vertic horizons. The Mg/Ca ratio is higher than 1 but rarely exceeds 2. Magnesium remains the dominant cation even when gypsum is present. The deep horizons are relatively richer in exchangeable sodium.

Hypermagnesic vertisols have a higher exchange capacity (50 and 60 me/100g soil), which might well be due to the higher clay contents and better crystallization of the smectites. The vertic horizons are base saturated. The Mg/Ca ratio is higher than 4.5 and increases with depth. The proportion of Mg++ in relation to the sum of exchangeable cations is higher than 80%. The deep horizons are also high in exchangeable sodium.

Sodic and acid vertisols have an exchange capacity which is relatively low in the surface horizons (15 to 20 me/100g soil) but increases a little with depth at the same time as the clay content, reaching 30 to 35 me/100g. The base saturation of the vertic horizon is slightly lower than that for other vertisols (80 to 90%). The exchangeable sodium content is often greater than 10 or more. The Mg/Ca ratio increases considerably

Figure 3. Amounts of exchangeable Ca++, MG++, and Na+ in the three main groups of Vertisols



from the surface (mean value of 2) toward the deeper horizons, where it often exceeds 5.

In summary, the following are very important in the characterization of the various types of vertisols in New Caledonia:

1. Magnesium is the dominant cation in the vertic horizons, including calcimagnesic vertisols with accumulations of calcium salts (gypsum, lime).
2. The exchangeable calcium content is relatively higher in the surface horizons than in the vertic horizons.
3. The exchangeable sodium content increases rapidly at depth in the vertic horizons. A considerable proportion of the sodium, in all probability, can be attributed to sea spray and aerosols.
4. The exchangeable potassium content is very low. Only rarely, however, is it completely absent.

Other Elements

The total phosphorus contents are very low, usually less than 800 ppm. The available phosphorus contents (Olsen method, modified by Dabin, 1967) are less than 20ppm. (Hypermagnesic vertisols developing from ultrabasic rocks are very rich in chromite, an element which masks phosphorus when the colorimetric method is applied. Use of the X-ray fluorescence method, however, also reveals very low phosphorus contents [Latham, 1986].)

Total analyses show a $Si0_2/Al_20_3$ molecular ratio which is always above 4. This ratio may reach much higher levels in vertisols derived from serpentinites, which are rocks totally lacking aluminium (hypermagnesic vertisols). The mean total iron contents mostly vary between 8 and 10% but reach 20% in the hypermagnesic vertisols. Of the alkaline and alkaline-earth elements, magnesium is in most cases the most abundant (with the exception of soils having developed from carbonate-rich flysches and limestones). Sodium is often more abundant than calcium. It should be noted that, in the horizons which show traces of manganese accumulation, this element, on average, reaches contents of 2 to 3% and even, in infrequent cases, 10%. The vertic horizons of the hypermagnesic vertisols which derive from peridotites and serpentinites are enriched in chromium, nickel, and cobalt.

Discussion and Conclusion

The data relating to exchangeable bases and the exchange capacity suggest two means of classification for vertisols in New Caledonia.

The first would be based on the exchangeable Mg^{++}/Ca^{++} ratio. Using this method we may distinguish between:

- calcic vertisols: the Mg^{++}/Ca^{++} ratio is less than 0.5 in the vertic horizons.
- calcimagnesic vertisols: the Mg^{++}/Ca^{++} ratio is between 0.5 and 2.0 in the surface horizon and rises to 3.0 in the vertic horizons.
- magnesic vertisols: the Mg^{++}/Ca^{++} ratio is between 3.0 and 5.0 in the surface horizon and rises to 10.0 in the vertic horizons.
- hypermagnesic vertisols: the Mg^{++}/Ca^{++} ratio is higher than 5.0 in the surface horizon and exceeds 10.0 in the vertic horizon.

The second method of classification is based on the ratio of exchangeable Na^+ percentage (ESP) measured in the vertic horizons (depth greater than 60 cm). It is thus possible to distinguish among:

- non sodic vertisols: ESP is less than 8.0
- moderately sodic vertisols: ESP is between 8.0 and 15.0.
- sodic vertisols: ESP is greater than 15.0.

It is thought that these vertisols were probably formed under the influence of a paleoclimate much dryer than the current climate, because the accumulations of gypsum are unlikely to have developed in the currently prevailing

weather conditions in New Caledonia.

The hypermagnesic vertisols are subject to the standard use constraints typical of all vertisols and also to those due to the major cationic imbalance of the exchange complex which is marked by the distinct predominance of magnesium. This magnesium surplus leads to calcium deficiencies. Amendments in the form of gypsum make it possible to restore the balance in these vertisols (Beaudou et al., 1984) and substantially increase yields (Bonzon, 1986, 1988).

Literature Cited

A.F.E.S., 1988. Référentiel Pédologique Français, 2nd Ed., INRA, 251 p.

Beaudou, A.G., M. Fromaget, P. Podowojewski, E. Bourdon, H. Le Martret, and D. Blavet. 1983. Cartographie typologique des sols: méthodologie, Orstom, Nouméa, 31 p.

Beaudou, A.G., H. Le Martret, and B. Denis, 1984. Magnesian soils of New Caledonia. Experiment to restore the balance of the base exchange complex. South Pacific J. of Natural Sc., vol.7, 78-99.

Bonzon, B., et al. , 1986. Effet des amendements calciques sur un sol sodique acide et sur un vertisol hypermagnésien, Orstom, Nouméa, 36 p.

Bonzon, B. et al. , 1988. Effet des amendements calciques sur vertisol hypermagnésien, Orstom, Nouméa, 69 p.

Dabin, B., 1967. Application des dosages automatiques à l'analyse des sols. Cah. Orstom, sér. Pédol., vol. V, no. 3.

Denis, B., and P. Mercky, 1982. Etude pédologique de la région de Pouembout, Orstom, Nouméa, Tome 1:150 p., Tome 2:109 p.

Latham, M., P. Quantin, and G. Aubert, 1978. Etude des sols de Nouvelle Calédonie, Orstom, Paris, Not. Expl. no. 78, 138 p.

Latham, M. 1986. Altération et pédogenèse sur roches ultrabasiques en Nouvelle Calédonie, Orstom, Paris, Et. & Thèses, 331 p.

Podowojewski, P. 1984. Les sols de Nouvelle Calédonie à accumulation de Gypse, Orstom, Nouméa, 23 p.

Podowojewski, P. and A.G. Beaudou, 1987. Carte morphopédologique de la Nouvelle Calédonie à 1/200.000, Orstom, Nouméa, 3t.

Tercinier, G. 1953. Sols et Terres de la Nouvelle Calédonie et autres territoires Français du Pacifique. 8ème congrès Pac. Sci., Manille, 1953.

Tucker, B.M. 1985 a. Active and exchangeable cations in soils, Aust. J. Soil Res. 23, 195-201.

Tucker, B.M. 1985 b. Laboratory procedures for soluble salts and exchangeable cations in soils, CSIRO, Soils Techn. Pap., 47 p.

Salinity Development, Recognition, and Management in North Dakota Soils

J.L. Richardson[1], D.G. Hopkins, B.E. Seelig, and M.D. Sweeney

Abstract

The field recognition of salinity is complicated because soluble salts are expressed morphologically only when the soil is very dry. Salinity can be assessed directly by morphological observation. Vegetation indicators are used, but they introduce a degree of unreliability. Plant indicators vary because of moisture conditions, time of year, and plant species. A significant problem in both soil survey and subsequent management of these saline and sodic areas is recognizing the intensity of the problem. In sodic map units, recognition of soil variability is problematic because of the scarcity of soil chemical data and variations in landscape expression in different sodic map units. In western ND more saline and sodic soils occur on terraces and floodplains. Also saline seeps on erosional footslopes are important. In eastern ND the salinity is associated with pond edges in Calciaquolls. A district system of evaporite sequences occur from calcite, then gypsum and lastly efflorescent crusts. Freezing also concentrates salt and changes the salt type. Most salts in ND are sulfates; on freezing $Na_2SO_4H_2O$ precipitates more easily than $MgSO_4H_2O$.

Introduction

North Dakota has two distinct geologic regions: 1) Quaternary terrains of till, outwash, and lacustrine sediments; and 2) Tertiary sediments over Cretaceous sands and shales. The Quaternary deposits occur often as closed drainage systems. They do not have outlets; we often call these areas "prairie potholes." Salinity associated with closed or partially closed drainage systems occur mostly in three major physiographic provinces in North Dakota: 1) Drift Prairie; 2) Missouri du Coteau; and 3) Glacial Lake Agassiz. Prairie potholes have been drained extensively throughout the region with open ditches constructed as outlets for waterflow. Salinity problems are abundant because of the youthful nature of the landscape and shallow water tables. Many of the saline soils are Calciaquolls that are essentially soils with evaporite deposits that have formed Bk horizons.

In the southwestern part of the state, the residual soils formed in bedrock sediments that have open-erosional landscapes with abundant streams and few natural wetlands. Salinity is associated with sodicity in footslopes and toeslopes. Due to the practice of summer fallowing and the stratified nature of the Tertiary sediments, saline seeps are common.

[1]Dep. Soil Science, North Dakota State Univ., Fargo, ND 58105. Published with the permission of the Director of the North Dakota Agric. Exp. Sta. Corresponding author.

This paper outlines salinity occurrence in North Dakota by region and discusses management and recognition of these soils.

Recognition of Salinity for Soil Survey

The exception is gypsum, which precipitates from solution at concentrations of about 2 g/L^{-1} or 2 dS/m^{-1}. At 20° C sodium sulfate precipitates approximately at 200 g/L^{-1} or > 200 dS/m^{-1}, magnesium sulfate at > 300 g/L^{-1} or > 300 dS/m^{-1}, and sodium chloride at > 375 g/L^{-1} or > 375 dS/m^{-1} (Timpson et al, 1986). Such extreme salinity levels are reached only in surface crusts or in rare circumstances.

In recent years, electromagnetic induction devices have improved our confidence in delineating saline soils (Richardson and Patterson, 1986; Wollenhaupt et al., 1986).

In North Dakota, strong (> 16 dS/m^{-1}) salinity is recognized at the series level. Slight and moderate salinities (4 to 16 dS/m^{-1}) are recognized as saline phases. For example, in Grand Forks County (Doolittle et al, 1981), Bearden, Bearden saline, and Ojata represent increasing salinity from very slight to strong. The members of the soil survey team in this county were uncertain about saline soil mapping unit delineations, but examination of the field delineations demonstrated that the units were accurate (Richardson and Patterson, 1986). The delineations were tested with an electromagnetic induction meter, the Geonics EM-58 that has been

used throughout the northern Great Plains (Wollenhaupt et al., 1988).

Hopkins et al. (1987) found forage yield estimates on sodic rangeland were vastly underestimated because a Leptic Natriboroll was overmapped due to the striking visual impact of barren panspots on the soil landscape.

Recent work in North Dakota has shown that saline seep formation resulting from agricultural management is actually a form of desertification (Timpson and Richardson, 1986).

Classification of Saline and Sodic Soils

Sodic soils with significant accumulations of soluble salts are classified as saline-sodic (Salinity Lab Staff, 1954). To meet the saline-sodic criteria, the soil saturation extract must have an electrical conductivity (EC) > 4 dS/m^{-1} and soil exchangeable sodium percentage (ESP) > 15. Saline-sodic soils are often flocculated due to high soil solution EC and may physically resemble saline soils. Some saline Calciaquolls meet the requirements for saline-sodic soils but unfortunately have not been separated from saline Calciaquolls. Sodic soils by contrast have morphological differences due to clay dispersion by sodium. Some sodic soils with high amounts of dispersible clay, such as Leptic Natriborolls and Typic Natraquolls, also meet the criteria for saline-sodic soils. In North Dakota a different definition for sodic-saline soils has been used. Saline-sodic soils must have a *dense subsoil* accompanied by a saturation extract EC greater than 2 dS/m^{-1} and ESP greater than 15.

Recognition and delineation of saline-sodic soils is complicated by variability in sodicity and salinity. Seelig and Richardson (1989) found dispersible clay differences between sodic soils to be related to landscape position. We found that identification of landscape position and soil moisture regime was necessary to predict saline-sodic properties.

Landscape and Saline Relationships

Salinity In Erosional Topographies Of Western North Dakota

Of fundamental concern is the recognition that distribution of saline soils is governed by landscape position. Saline or sodic soils develop in open or closed geomorphic systems wherever convergent water flow concentrates salts. Thus soils in a drainage basin are related by the transfer of chemical constituents and can be thought of as representing a "single geochemical landscape" as pointed out by Huggett (1975). Recently, research emphasis has shifted to the accelerated rate of secondary soil salinization caused by cultural management (Halvorson and Black, 1974; Doering and Sandoval, 1978; Ferguson and Bateridge, 1982; Timpson and Richardson, 1986).

In western North Dakota, saline and sodic soils occur in three distinct landscape positions. The largest areas of saline and sodic soils are found in broad flats of floodplains and terraces. Saline groundwater is often shallow and moves upward into the soil root zone by capillary rise. Saline phases of Torrifluvents and Ustifluvents and the Natrargids, Natraquolls, and Natriborolls are found in these landscapes.

Saline seeps frequently occur on upper pediment footslopes. Here, small scale stratigraphic controls result in seepage at the slope break associated with the backslope. Panspots, a ubiquitous surface feature found in sodic landscapes, are related by a similar process. They generally merge with adjacent upland soils at a scarp face resembling a micro-pediment backslope. The higher salinity in the panspot soils is likely the result of throughflow seepage caused by structural and/or textural control on permeability in the adjacent upland soils (Murphey and Daniel, 1935; Bowser et al., 1962; Hopkins et al., 1987).

The most extensive landscape position is on lower erosional footslope positions. The majority of Natrargid and Natriboroll soils are found in these landscape positions (Rosek and Richardson, 1989) over extensive areas. Seasonal throughflow is vertically controlled above relatively impermeable substratum and converges in lower footslope positions either as perched shallow groundwater or as seepage (Hadley and Rolfe, 1955).

Salinity In Till And Lacustrine Landscapes

Numerous wetlands and wet soils result from the lack of an integrated stream flow network in the Prairie Pothole Region and in the Lake Agassiz Physiographic Province. Groundwater movement (Miller et al., 1985; Arndt and Richardson, 1988) controls many of the processes of soil development. Therefore, recognition of the pathways of water movement are important in understanding soil development

and in particular salinity (Bigler and Richardson, 1984; Richardson and Bigler, 1984).

Several studies have helped our understanding of the groundwater movement in these systems. Lissey (1971), in particular, introduced the notion of depression-focused recharge and discharge (Fig. 1). In these areas, water flows to the groundwater only in the smaller depressions (recharge). Water flows from groundwater to the surface in lower wetlands (discharge). The wetlands or prairie potholes are "windows" of the groundwater at the earth's surface. The discharge points become progressively more saline. Note that the water tends to flow laterally in the groundwater system as suggested by Lissey (1971). Such water flow systems are expected in the subhumid and drier climate zones.

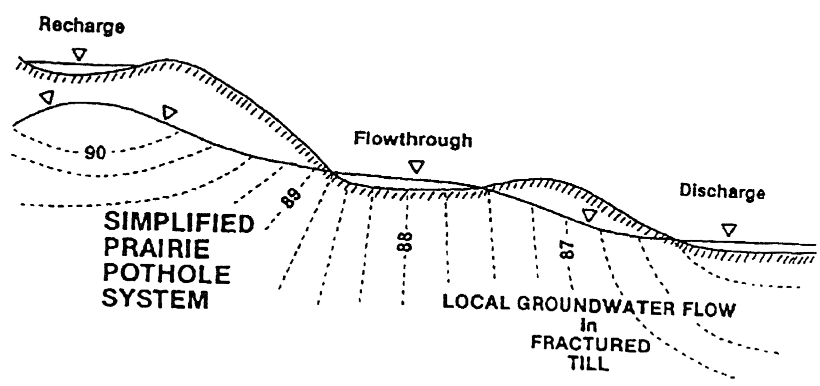


Figure 1. A simplified flownet with equipotential lines in meters for the Prairie Pothole region. Note that water table is mounded under wetlands and does not follow the topography well (Modified from Lissez, 1971).

MacLean and Pawluk (1975) observed lateral flow in relatively open systems and correlated salinity and groundwater. They also observed desalinized-sodic systems that were apparently due to lateral flow above the water table. Mills and Zwarich (1986) and Winter (1986) observed lateral flow in a variety of prairie pothole landscapes. They noted many seasonal flow reversals complicating the flow system.

Salinity As Evaporite Deposits Associated With Wetlands

Knuteson et al. (1989) noted that depression-focused recharge affected soil development in lacustrine landscapes. A Calciaquoll is expected to form around ephemeral wetlands, but leached soils occur in them. Upward movement of capillary is dominant water at the pond edge, even though saturated flow may reverse around many ponds. Unless a groundwater discharge area is encountered, the salt content remains low. Calcite is the main evaporite concentrated in the Calciaquoll soils on the margin around recharge wetlands. Saline soils occur on the margins of discharge wetlands that have concentrations of evaporites more soluble than calcite (Steinwand and Richardson, 1989; Arndt and Richardson, 1989).

The combination of discharge and evapo-transpiration create an "edge effect" in which evaporites accumulate (Fig. 2). This observation was earlier noted by Whittig and Janitzky (1963) in California. They found saline and sodic conditions similar to those in North Dakota wetlands, but their warmer environment with low sulfates produced an evaporite sequence dominated by Na_2CO_3. In North Dakota, the observation of Na_2CO_3 is unusual. Abundant sulfate salts probably result from oxidation of reduced sulfur in organic matter or pyrites that were originally in marine shales (Hendry et al., 1986; Mermut and Arshad, 1987).

The Hardie and Eugster (1970) system of closed basin brine evolution works well to explain the distribution and type of salts in saline soils from North Dakota (Skarie et al., 1986, 1987; Arndt and Richardson, 1989) (Fig. 3). In North Dakota, sulfatic evaporites are the dominant source of salinity. The first evaporites to form are magnesium calcites in the northern Great Plains (St. Arnaud and Herbillon, 1973; Knuteson, 1985).

With a binary salt, such as calcite, either the anion or the cation phase will increase in the soil solution by the precipitation of the salt. In the till soils, continued precipitation of calcite caused by evapotranspiration concentrates dissolved Ca^{2+} and depletes soluble CO_3^{2-}. This is an illustration of Hardie and Eugster's (1970) chemical divides. Note that in outwash areas HCO_3^{2-} was observed to be more abundant than Ca^{2+}, and the pH usually exceeded 8.4. In warmer climates soil solutions would not hold the CO_2 in solution as well as in North Dakota. This may be another reason for the differences between North Dakota wetland edges and the edges observed by Whittig and Janitzky (1963) in California.

Last and Schweyen (1983) noted several alkaline lakes with high carbonate. In these lakes the evaporites contained protodolomite. In one alkaline soil in North Dakota (maximum pH 9.3), dolomite was observed as well. Two other cases of dolomite in soils of the northern Great Plains have been reported (Sherman et al., 1962; Rostad, 1975). Therefore, it appears that, in soils with more CO_3^{2-} than Ca^{2+}, the precipitation of calcite leads to high pH levels and dolomite. The conditions of dolomite formation require very high levels of Mg^{2+} with respect to Ca^{2+} as well as elevated pH levels. This is one of the "chemical divides" of Hardie and Eugster (1970).

Keller et al. (1986) observed that a few salt efflorescent evaporites contained burkeite ($Na_6(CO_3)_2SO_4$) and tychite ($Na_6Mg_2(CO_3)_4SO_4$). These are "alkaline" evaporites and would be the equivalent of the sodium carbonates of Whittig and Janitzky (1963). The alkaline evaporites have been observed to occur only in coarse textured soils; rapid degassing of CO_2 may contribute to their formation.

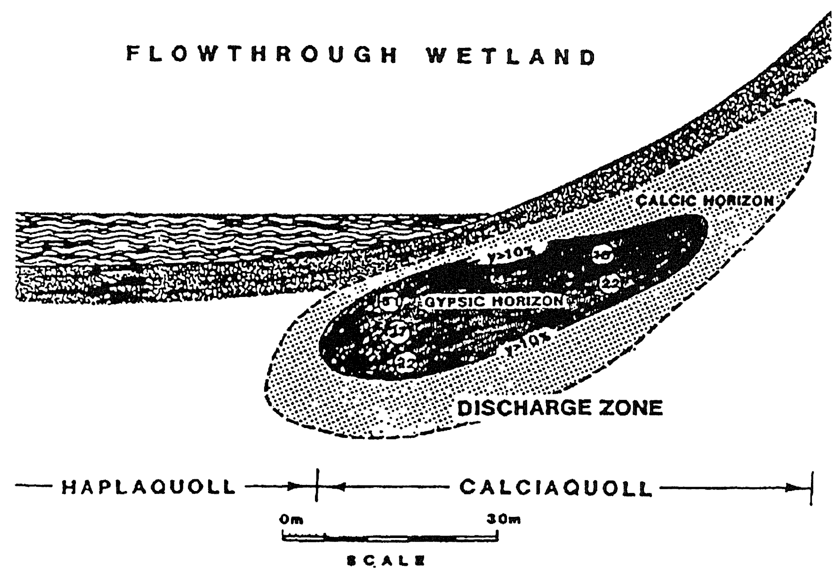


Figure 2. An example of the evaporite deposition at pond edges in the Prairie Pothole Region. Numbers are for gypsum by weight. Average EC was > 16 mmhos/cm (after Steinwand and Richardson, 1989).

The more typical evaporite condition, however, contains more Ca^{2+} than HCO_3^- (Skarie et al. 1987; Arndt and Richardson, 1989; Steinwand and Richardson, 1989). Gypsum is the mineral that has been observed to precipitate under these conditions; the Hardie and Eugster model predicts this observation. The ubiquitous nature of gypsum in saline and sodic soils of North Dakota indicates that the soil solution is saturated with respect to gypsum. It also indicates that gypsum is an ineffectual reclamation material in high sodic saline soils. As gypsum forms, the Ca^{2+} is depleted from the soil solution (Skarie et al., 1987 and Arndt and Richardson, 1989). The net effect is that sodium and magnesium sulfates concentrate in the soil solution. These salts are generally the most abundant saline soils from North Dakota. They usually precipitate only in soil surfaces in highly concentrated forms as efflorescent crusts (Keller et al., 1986).

Most saline soils from North Dakota do not have precipitates of soluble "salts." Frequently soil surveyors observe "salts" *in* the soil as opposed to *on* the soil. These salts are most likely gypsum and may or may not indicate saline conditions.

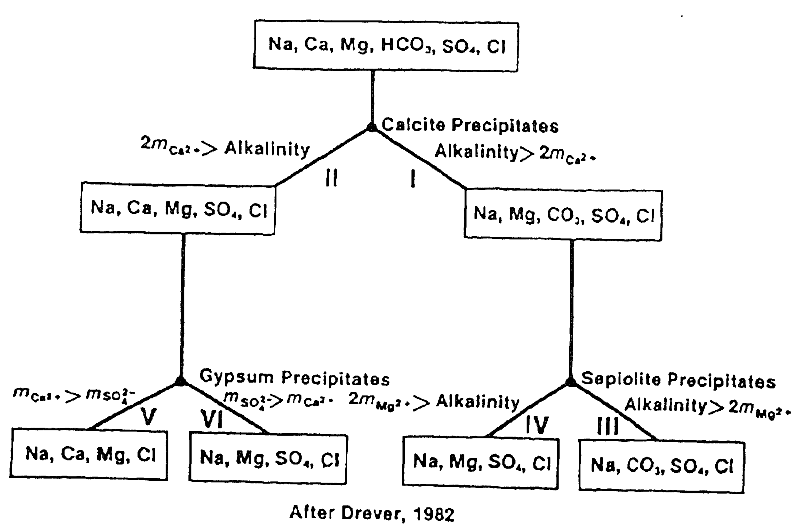


Figure 3. Hardie and Eugster (1970) chemical divides model for closed basin brine evolution. The evaporites of the Prairie Pothole conform well to the model even though our system is open.

Salinity and Freezing

Another factor in northern soils is the effect of freezing on salt concentration. Timpson et al (1986)

and Keller et al (1986) noted the temperature change effects on sulfate minerals. Arndt and Richardson (1986, 1989) observed that ice on lakes contained the sodium sulfate mineral, mirabilite, and the lake water was concentrated with magnesium sulfate. Beke and Palmer (1989) observed mirabilite in frozen soils of southern Alberta.

Freezing can segregate sodium from magnesium in sulfate rich saline soils. This may increase Mg^{2+} on the exchange complex in the fall and winter. On warming, a flush of ice-freed sodium will dissolve into the soil solution and can temporarily create a mass action exchange with soils. The freezing action in combination with ion pairing (Alzubaidi and Webster, 1983) should maximize the influence of sodium. In such conditions "magnesium solonetzs" can be created with a minimum of sodium.

Management

Management Of Dryland Saline Soils

Dryland saline soils are the best managed as range land (pasture land) or hay land. Tillage concentrates salts and creates surface efflorescences that only partially are revegetated, if at all. Caution is advised with regard to drainage. Skarie et al. (1987) noted that drainage in many flat landscapes increased salinity along the ditches. Griffin et al. (1985) also found that lagoons or any other source of water in these landscapes raised the local water table and caused saline soils.

Richardson and Arndt (1989) have noted that any wetland that receives groundwater would likely be saline if drained. It is suggested that any wetland with a component of discharge should not be drained; this includes all semipermanent ponds. Further, a substantial zone of native grasses should be left around the wetland. Salinization of soils adjacent to wetlands may occur by plowing too close to a pond edge.

It is clear that selected grazing systems can increase total range production for livestock and wildlife on soils that are prone to salinization. Sedivec (1989) noted that grazing is a compatible land use with waterfowl production.

Saline seeps are common in the southwestern part of the state and on the Missouri Coteau. These are emphemeral springs of saline water that appear during periods of higher than usual precipitation. Ferguson and Bateridge (1982) found that crop-fallow management increased deep percolation of water that contributed to lateral subsurface flow to saline seeps. Saline seeps are best managed by reducing deep percolation of water in the recharge areas by planting crops. We recommend a management system that includes deep rooted crops, such as alfalfa, and no summer fallow. Division ditching and tiling removes water from the seep directly.

Management Of Saline-Sodic Soils

The first rule to note in the northern Great Plains is that reclamation of a natric or sodic condition with gypsum will not be successful in most cases. The soil solution is saturated with respect to gypsum; and adding more gypsum will not make it any more soluble. Calcium chloride is suggested as the amendment of choice in sulfatic soils.

Management of saline-sodic soils depends not only on chemical and physical soil properties, but also on soil moisture regime. Soluble salt removal and replacement of exchangeable sodium by calcium will reclaim these soils (Salinity Lab Staff, 1954), but the moisture regime must allow adequate leaching. Often this is not the case, because these soils have formed due to high water tables.

Seelig and Richardson (1989) found that the most severely affected sodic soils had the highest salinity. Decisions to improve the moisture regime by drainage should be made judiciously. Attempts to drain and leach saline-sodic Calciaquolls are likely to lead to a puddled soil due to clay dispersion upon removal of the soluble salts (Salinity Lab Staff, 1954). Sandoval (1978) reported that deep plowing improved a saline-sodic Leptic Natriboroll in western North Dakota. This technique is most successful on the drier sodic soils.

The key to proper management of saline-sodic soils is accurate determination of soil moisture regime. Saline-sodic soils with drier moisture regimes are less restrictive to use and management than their wetter counterparts and have the greatest reclamation potential.

Management Of Irrigated Soils

Salinity problems in the western states occur from the improper application of irrigation water. Numerous soils have become saline through the use of poor quality irrigation water, and, in many cases, lack of adequate leaching has complicated the management of certain soils (Sweeney, 1973).

Drainage is an important requisite, because excess water in the root zone restricts root

growth, delays warming of the soil in the spring, and induces the accumulation of salts. Subsurface drains often are required to maintain the water table below the rooting depth of the crop and to provide for the removal of additional water necessary for leaching of excess soluble salts.

High concentrations of salt in the soil increase the energy expended by the plant to obtain water and reduce the evapotranspiration and growth rates of the plant. It increases the irrigation requirement, because additional water in excess of that needed for evapotranspiration must be applied to leach the salts from the root zone. Most saline soils can be improved by leaching with good quality water because many of the salts are very soluble. Soil physical conditions for crop production are generally good.

To prevent secondary salinization by irrigation, an inventory of soils' water quality should be made. The need for surface and/or subsurface drains, soil amendments, deep tillage, and salt tolerant crops can be assessed. Proper management based on the needs assessment will minimize the potential for salinization.

References

Alzubaidi, A., and G.R. Webster. 1983. Ion pairs in a solonetzic soil Can. J. Soil Sci. 63:479-484.

Arndt, J.L., and J.L. Richardson. 1986. The effects of groundwater hydrology on salinity in a recharge-flowthrough-discharge wetland system in North Dakota. p. 269-277. In G. Van der Kamp and M. Madunicky (eds.) Proceedings of the Third Canadian Hydrogeological Conference, Saskatoon Sask., 21-23 April 1986.

Arndt, J.L., and J.L. Richardson. 1988. Hydrology, salinity and hydric soil development in a North Dakota prairie-pothole wetland system. Wetlands 8:93-108.

Arndt, J.L., and J.L. Richardson. 1989. Geochemical development of hydrogen soil salinity in a North Dakota prairie-pothole wetland system. Soil Sci. Soc. Am. J. 53:848-855.

Beke, G.J. and C.J. Palmer. 1989. Subsurface occurrence of mirabilite in a Mollisol of southern Alberts, Canada: A case study. Soil Sci. Soc. Am. J. 53:1611-1614.

Bigler, R.J., and J.L. Richardson. 1984. Classification of soils in prairie pothole wetlands with deep marsh plant species in North Dakota. Soil Survey Horiz. 25:16-24.

Bowser, W.E., R.A. Milne, and R.R. Cairns. 1962. Characteristics of the major soil groups in an area dominated by Solonetizic soils. Can. J. Soil Sci. 42:165-179.

Doering, E.J. and F.M. Sandoval. 1978. Chemistry of seep drainage in southwestern North Dakota. Proc. Subcommission on Salt Affected Soils. 11th Intern. Congr. Soil Sci., Edmonton, Alberta., Vol 3 pp 1-15.

Doolittle, J.A., C.J. Heidt, S.J. Larson, T.P. Ryterske, M.G. Ulmer, and P.E. Wellman. 1981. Soil survey of Grand Forks County, North Dakota. U.S. Department of Agriculture-Soil Conservation Service, U.S. Government Printing Office, Washington DC, U.S.A.

Eugster, H.P., and B.F. Jones. 1979. Behavior of major solutes during closed-basin brine evolution. Am. J. Sci. 279:609-631.

Ferguson, H., and T. Bateridge. 1982. Salt status of glacial till soils of north central Montana as affected by the crop-fallow system of dryland farming. Soil Sci. Soc. Am. J. 46:807-810.

Griffin, D.M. Jr., R.L. Skarie, A. Maianu, and J.L. Richardson. 1985. Effects of prolonged lagoon leakage on agricultural land. J. Civil Engr. 4:797-806.

Hadley, R.F. and B.N. Rolfe. 1955. Development and significance of seepage steps in slope erosion. Trans. Am. Geophys. Union 36:792-804.

Halvorson, A.D. and A. L. Black. 1974. Saline seep development in dryland soils of northwestern Montana. Soil and Water Cons. J. 29:77-81.

Hardie, L.A., and H.P. Eugster. 1970. The evolution of closed-basin brines. Miner. Soc. Am. Spec. Paper 3:273-290.

Hendry, M.J., J.A. Cherry, and E.I. Wallick. 1986. Origin and distribution of sulfate in a fractured till in southern Alberta, Canada. Water Resources Research 22:45-61.

Hopkins, D.G., J.L. Richardson, and M.D. Sweeney. 1987. Composition comparisons in sodic map unit delineations on the Dickinson Experimental Station Ranch Headquarters, North Dakota. Soil Survey Horiz. 28(2):46-50.

Huggett, R.J. 1975. Soil landscape systems: a model of soil genesis. Geoderma 13:1-22.

Keller, L.P., G.J. McCarthy, and J.L. Richardson. 1986. Mineralogy and stability of soil evaporites in North Dakota. Soil Sci. Soc. Am. J. 50:1069-1071.

Knuteson, J.A. 1985. Microrelief and pedogenesis of soils of the Lake Agassiz plain. Ph.D. Diss., North Dakota State Univ., Fargo. (Diss. Abstr. 86-06139).

Knuteson, J.A., J.L. Richardson, D.D. Patterson, and Lyle Prunty. 1989. Pedogenic carbonates in a Calciaquoll associated with a recharge wetland. Soil Sci. Soc. Am. J. 53:495-499.

Last, W.M., and T.H. Schweyan. 1983. Sedimentology and geochemistry of saline lakes of the Great Plains. Hydrobiologia. 105:245-263.

Lissey, A. 1971. Depression-focused transient ground water flow patterns in Manitoba. Geological Association of Canada Special Paper 9:333-341.

MacLean, A.H., and S. Pawluk. 1975. Soil genesis in relation to groundwater and soil moisture regimes near Vegreville, Alberta. J. Soil Sci. 26:278-293.

Miller, J.J., D.F. Acton, and R.J. St. Arnaud. 1985. The effect of ground water on soil formation in a morainal landscape in Saskatchewan. Can. J. Soil Sci. 65:293-307.

Mills, J.G., and Zwarich. 1986. Transient ground-water flow surrounding a recharge slough in a till plain. Can. J. Soil Sci. 66:121-134.

Mermut, A.R., and M.A. Arshad. 1987. Significance of sulfide oxidation in soil salinization in southeastern Saskatchewan, Canada. Soil Sci. Soc. Am. J. 51:247-251.

Murphy, H.F. and D.A. Daniel. 1935. Some chemical and physical properties of normal and Solonetz soils and their relation to erosion. Soil Sci. 39:453-461.

Richardson, J.L., and R.J. Bigler. 1984. Principal component analysis of prairie pothole soils in North Dakota. Soil Sci. Soc. Am. J. 48:1350-1355.

Richardson, J.L., and D.D. Patterson. 1986. Verification of saline mapping units for a soil survey. Soil Survey Horiz. 27:22-27.

Richardson, J.L., and J.L. Arndt. 1989. What use prairie potholes. Soil and Water Conserv. J. 44:No. 3:196-198.

Rosek, M.J., and J.L. Richardson. 1989. Comparison of erosional footslope soils in high and low relief drainage basins of western North Dakota. USA. Geoderma (Accepted for Publication).

Rostad, H.P.W. 1975. Diagenesis of postglacial carbonate deposits in Saskatchewan. Can. J. Earth Sci. 12:798-806.

St. Arnaud, R. J. and A.J. Herbillon. 1973. Occurrence and genesis of secondary magnesium-bearing calcites in soils. Geoderma 9:279-298.

Salinity Laboratory Staff. 1954. Diagnosis and improvement of saline and alkali soils. Agric. Handb. No. 60, USDA. U. S. Gov. Print. Off. Washington, D. C.

Sandoval, F.M. 1978. Deep plowing improves sodic claypan soils. North Dakota Farm Res. 35:4:15-18.

Sedivec, Kevin. 1989. Effects of specialized grazing systems on upland nesting waterfowl production in southcentral North Dakota. Master of Science Thesis. Animal and Range Sci. Dep., North Dakota State Univ., Fargo.

Seelig, B. D, and J.L. Richardson. 1989. Soil profile characteristics on a sodium affected landscape in central North Dakota. Papers presented at the 32nd Annual Manitoba Society of Soil Science Meeting. Dep Soil Sci., Univ. Manitoba, Winnipeg. pp 46-61.

Sherman, G.D., F. Schultz, and F.J. Alway. 1962. Dolomitization in soils of the Red River Valley, Minnesota. Soil Sci. 94:304-313.

Skarie, R.L., J.L. Richardson, A. Maianu, and G.K. Clambey. 1986. Soil and groundwater salinity along drainage ditches in eastern North Dakota. J. Environ. Qual. 15:335-340.

Skarie, R.L., J.L. Richardson, G.J. McCarthy, and A. Maianu. 1987. Evaporite mineralogy and groundwater chemistry associated with saline soils in eastern North Dakota. Soil Sci. Soc. Am. J. 51:1372-1377.

Steinwand, A.L., and J.L. Richardson. 1989. Gypsum occurrence in soils on the margin of semipermanent prairie pothole wetlands. Soil Sci. Soc. Am. J. 53:838-842.

Sweeney, M.D. 1973. Soil and water characteristics important in irrigation. Circular S&F-573, Extension Service, North Dakota State University, Fargo.

Timpson, M.E., J.L. Richardson, L.D. Keller, and G.J. McCarthy. 1986. Evaporite mineralogy associated with saline seeps in southwestern North Dakota. Soil Science Society of America Journal 50:490-493.

Timpson, M.E., and J.L. Richardson. 1986. Ionic composition and distribution in saline seeps of southwestern North Dakota, USA. Geoderma 25:295-305.

Whittig, L.D., and P. Janitzky. 1963. Mechanism of formation of sodium carbonate in soil: I. Manifestation of biological conversion. J. Soil Sci. 14:322-333.

Winter, T.C. 1986. Effect of ground-water recharge on configuration of the water table beneath sand dunes and on seepage in lakes in the sandhills of Nebraska, U.S.A. J. Hydrology 86:221-237.

Wollenhaupt, N.C., J.L. Richardson, J.E. Foss, and E.C. Doll. 1986. A rapid method for estimating soil salinity from apparent soil electrical conductivity measured with an above-ground electromagnetic induction meter. Can. J. Soil Sci. 66:315-321.

Aridisols of Argentina

C.O. Scoppa and R.M. di Giacomo[1]

Abstract

The aridic soil moisture regime (SMR) covers some 1,670,000 km2, which is 59% of the total surface of Argentina (2,800,000 km2). It extends on a line crossing the country in a north-south direction and on which some 502,000 km2 of Aridisols (most of them Argids) have developed, representing 32% of the surface with an aridic SMR. The remaining 68% is covered by Entisols and some ustic, xeric, and aridic Mollisols, ustic Alfisols, and Inceptisols in the boundary areas with the Aridic regime.

The parent material of the Aridisols corresponds to sandy aeolian and fluvial sediments of Holocene age which are burying, at different depths, clayey sedimentary strata of Pliopleistocene age laying on different Prequaternarian rocks. This pattern of sedimentological stratification is repeated along the whole country, showing evidence of the action of exogen agents closely related to the relief.

The geographical distribution of the Aridisols, at a regional level, shows that they are rather common in the north (between 22 and 28 degrees south latitude) in a mountainous landscape with closed depressions where the Argids (Paleargids) and Orthids (Camborthids) are found in similar proportions. They are scarce in the planes of the great central sedimentary basin (between 28 and 39 degrees south latitude) where Orthids (Paleorthids, Calciorthids, and Camborthids) are dominant.

The most representative area is the Patagonia region, located in the southern portion of the country (between 39 and 52 degrees south latitude). The landscape is one of dissected tables where Argids and the great groups Natrargids, Paleargids, and Haplargids dominate over Calciorthids, Camborthids, and Paleorthids.

The areal distribution and the aforementioned taxonomic difference is due to the distribution and thickness of the Holocene sediments. This is a result of the prevailing winds and is closely related to relief and vegetation. These cold, dry, and intense winds originate at the South Pacific anticyclone. They have irregularly distributed the Holocene sedimentary layer, leaving at different depths the clayey Pliopleistocene materials, which are parent material of the argillic horizon of the Argids.

These Pliopleistocene materials had their origin under climatic and environmental conditions quite different from the present.

The difference at the great group and subgroup levels is given by the textural, physical, chemical, and physico-chemical features of the parent materials in combination or not with the present pedogenetic conditions.

Environmental Conditions of Argentina

Argentina is located between 66 degrees 57 minutes and 73 degrees 29 minutes W longitude and 21 degrees 46 minutes and 55 degrees 21 minutes S latitude. The surface is 2,800,000 km2.

The country offers different and contrasting landscapes of different origin, nature, and morphology: mountains and basins in the west, plains in the northeast and southeast, and tablelands in the south and northeast. The Cordillera de los Andes is the axis of these environmentally contrasting conditions, and the climate - in a NE-SW direction - varies from subtropical (humid) to temperate warm (humid, semi-arid, and arid) and temperate cold (arid and humid). According to the climate, the vegetation varies from subtropical forest, to park and subtropical savanna, pasture, semi-arid woodland, shrub desertic steppe, and desert of altitude.

Some 59% of the total surface (1,670,000 km2) shows an arid condition. The arid and semi-arid regions are extended across a longitudinal line approximately limited by two mean annual rainfall lines of 300 mm, one located to the east and the other one to the west, along the Cordillera de los Andes. In this frame, Aridisols - representing 32% of the total soils with and aridic SMR. of the country - have developed.

[1]Centro de Investigaciones de Recursos Naturales (CIRN). Instituto Nacional de Tecnologia Agropecuaria (INTA). Las Cabañas y Los Reseros s/n - (1712) - Castelar - Argentina. Phone 01-621-0281. Telex 17519 INTA AR, FAX 54-111-1917.

Pedoenvironmental Conditions of the Argentine Arid Region

Pedoclimate

The soil moisture and temperature regimes (SMR and STR) of Argentina are shown in Figure 1, taken from Van Wambeke and Scoppa (1976).

The distribution and extent of the aridic regime correspond to the general conditions of the existing climate. This SMR coincides in general with the arid and semiarid regions delimited by the 300 mm mean annual rainfall line and is only interrupted in the north, by an inclusion of udic and ustic regimes due to specific conditions of the relief. The STR are mesic and thermic in the north, thermic and hyperthermic in the center, and mesic and thermic in the south.

Parent Material

In the region of aridic SMR, most of the Prequaternarian rocks are covered by semi-consolidated sediments more or less clayey and presenting different concentrations and shapes of calcium carbonate, gypsum, and sodium of the Low Quaternary (Pliopleistocene?). Overlaying these sediments there is a sandy aeolian layer (Holocene?) more or less rich in pyroclastic material. The thickness and distribution of this layer varies with latitude. In some sectors this layer has been and is being worked by wind action or is covered by modern sandy sediments (Recent) of aeolian or fluvial origin.

The sedimentary formations of the Pliopleistocene and Holocene as a whole constitute the parent material of the Aridisols.

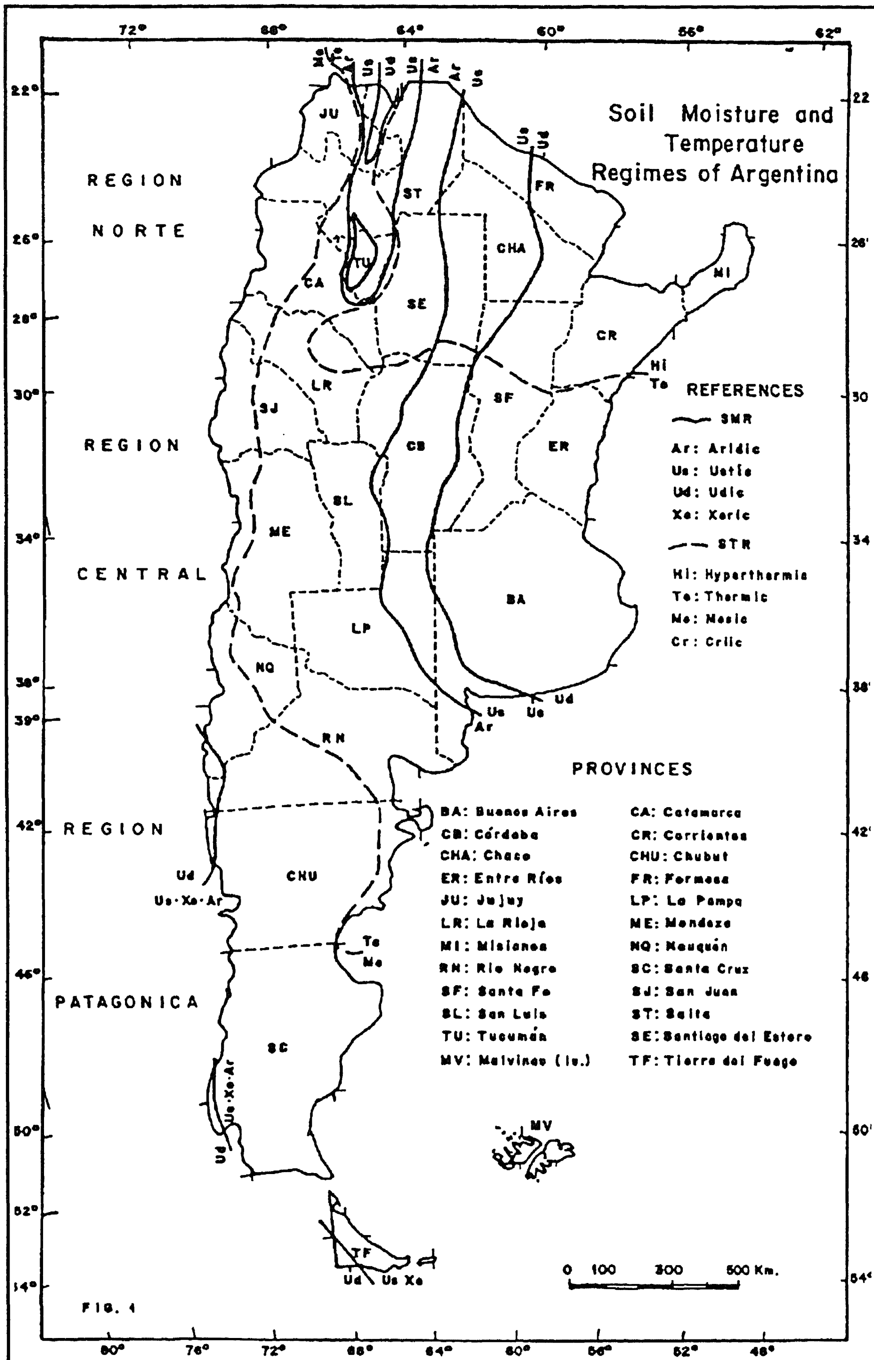


Figure 1

Relief

The relief is composed of three main elements: the mountains located in the west, following a north-south line along the country; the plains in the central and western parts; and the tablelands in the south (Figure 1). Most of the mountains correspond to the Cordillera de los Andes and other associated morphostructural elements. The height ranges from 4,000 to 7,000 m with peaks and valleys. In general the mountains follow a meridian line.

The plains extend east from the piedmont of the Cordillera, showing smooth shapes. In many cases they are of aeolian origin (dunes). The drainage pattern is very poor.

The tablelands correspond to dissected plains with stepped relief located east of the Patagonic

Cordillera and extending down to the Atlantic Ocean. The general characteristic of this region is the presence of a tabular relief falling to the river valleys or the sea. In some cases it is cut by low sierras, depressions, or basaltic plains.

Vegetation

The following vegetation patterns (modified from Cabrera, 1976) are found in the aridic moisture regime.

Woodland semideciduous xerophytic

It is found in the northeast and east central. The most common community is quebracho colorado (Schinapsis balansae), quebracho blanco (Aspidosperma quebracho blanco), algarrobo, and calden (Prosopis sp.). Gramineous savannas and halophilous steppes constitute secondary communities.

Shrub and gramineous steppe (altitude 2,000 - 4,500 m)

Located west of the preceeding. The characteristic feature is the absence of trees and the dominance of cactus, leguminous and composed shrubs.

Shrub and gramineous steppe (altitude 2,000 m)

A central strip extends from the boundary of the preceeding formation down to the Atlantic Ocean. Small tree species and zigophilaceous shrub of the Larrea sp. kind and Prosopis shrubs are dominant.

Shrub and Herbaceous steppe

Characteristic of the Patagonic region, it extends from the limit of the preceedingly mentioned formation down to the southern extreme of Argentina. The dominant species are neneo (Molinum spinosum), coirön amargo (Stipa sp.), coirön blanco (Festuca palescens), and Larrea sp.

TABLE 1: Taxonomy and inventory of Argentina Aridisols

SUBORDERS	GREAT GROUPS	SUBGROUPS
ARGIDS 306,396 Km2 61 %	NATRARGIDS 114,500 Km2 37 %	typic, aquic, ustollic, haplic, lithic,lithic-xerollic, borollic, xerollic and haploxerollic.
	PALEARGIDS 102,270 Km2 33 %	typic, petrocalcic, petrocalcic-xerollic, ustollic, xerollic and borollic.
	HAPLARGIDS 89,600 Km2 30 %	typic, borollic, aquic, arenic, lithic, lithic-xerollic, ustollic, xerollic,lithic-ustollic.
ORTHIDS 195,000 Km2 39 %	CALCIORTHIDS 79,600 Km2 41 %	typic, ustollic, lithic, lithic-ustollic, xerollic and borollic-lithic.
	CAMBORTHIDS 69,300 Km2 36 %	typic, borollic,fluventic, lithic, ustollic, lithic-xerollic and natric.
	PALEORTHIDS 32,400Km2 16%	typic,ustollic and xerollic.
	SALORTHIDS 11,000Km2 6%	typic and acuolic.
	GYPSIORTHIDS 2,000Km2 0,7%	typic, calcic and petrogypsic.
	DURORTHIDS 760Km2 0,3%	typic.

Taxonomy and Inventory

The Suborders, Great Groups, and Subgroups of Aridisols known in Argentina, as well as the surface they occupy and the percentage they represent up to the Great Group level, taken from Atlas de Suelos de la Rep#blica Argentina (1989), are shown in Table 1. These taxa are mostly associated with Entisols. Only in the border sectors of the Aridic regime, in transition to ustic and xeric regimes, can an association with Suborders, Great Groups, and Subgroups of ustic, xeric and aridic Mollisols, some Inceptisols, and ustic Alfisols be found.

Distribution

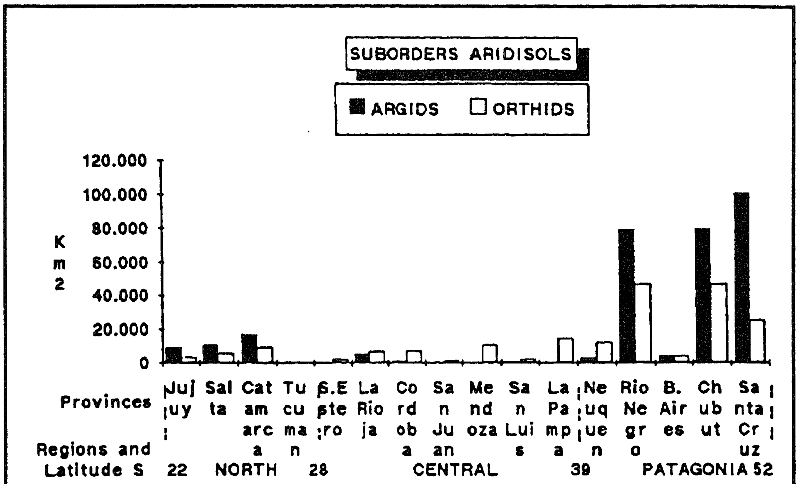
Within the great region of aridic SMR (1,670,000 km2), only 32% of this surface (502,000 km2) corresponds to Aridisols, distributed as follows: 306,389 km2 (61%) is occupied by Argids and 195,889 km2 (39%) by Orthids (Table 1).

The geographic distribution at the Suborder and Great Group levels is shown in Graphics 1, 2, and 3, disposed by provinces and regions from north to south, extended between 22 and 52 degrees south latitude.

If we analyze this distribution, it appears that most of the Aridisols (Argids + Orthids) of Argentina are in the Patagonic region (between 39 and 52 degrees South latitude) and that they are more or less equally represented in the central (between 28 and 39 degrees south latitude) and northern (between 22 and 28 degrees south latitude) regions.

At the Suborder level, Argids dominate in the Patagonic and northern regions, while in the central region Orthids are almost exclusive.

As for the Great Groups of the Argids, Paleargids are practically the only existing in the north. There are only a few Haplargids in the center, and Natrargids, Paleargids, and Haplar-



Graphic 1

gids are found in the south, with different proportions.

Within Orthids, Camborthids and Calciorthids along with very few Salorthids are the most common in the northern and central regions, whereas Calciorthids and Gypsiorthids are dominant in the Patagonia.

A summary of the above can be seen in Table 2. The percentage of each of the Suborders and Great Groups is given for every one of the three considered regions.

Characteristics of the Taxa in the Different Regions

A brief description of the associations, parent material, and landscape and some characteristics of the morphology of the Great Groups in the different regions considered is given below. In Table 3 a summary of the physical and chemical properties of these taxa is displayed.

Northern Region

Paleargids

These soils appear in associations of the subgroups: typic, ustollic, and petrocalcic and/or in association with Entisols. Paleargids have developed on aeolian friable materials - mixed with alluvial and colluvial materials - laying on rubble and gravel, on ancient stabilized forms: piedmonts, alluvial cones, and plains. The red argillic horizon often appears in the surface as a consequence of the erosion of the A1 horizon. A petrocalcic horizon can be present, separating the profile either from the Paleozoic rock or from Cenozoic sediments.

Haplargids

Associated with Entisols (Torriorthents) they also have developed on colluvial, alluvial, and aeolian sediments within natural drainage ways. The typic subgroup is found on fluvial terraces. The borollic subgroup appears in smooth relief areas with gentle slopes (Puna and Prepuna) and is above 4,000 m in elevation.

Natrargids

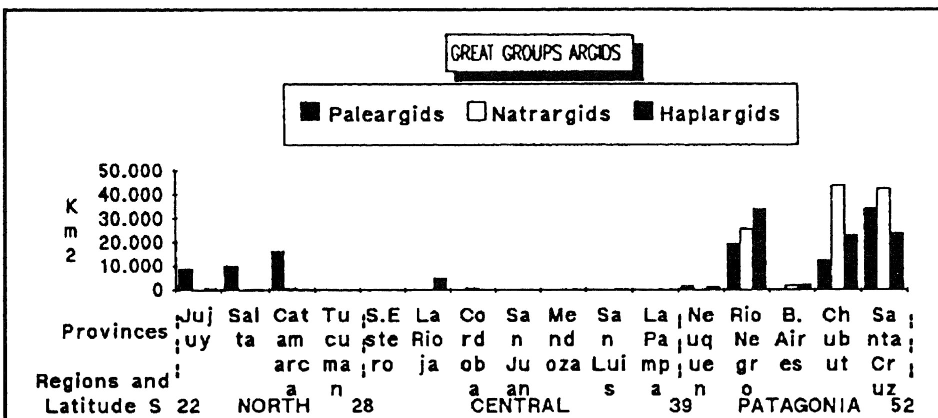
Associated with Entisols and other Argids, they have developed mainly in alluvial on fluvial plains and on flood plains with depressions.

Camborthids

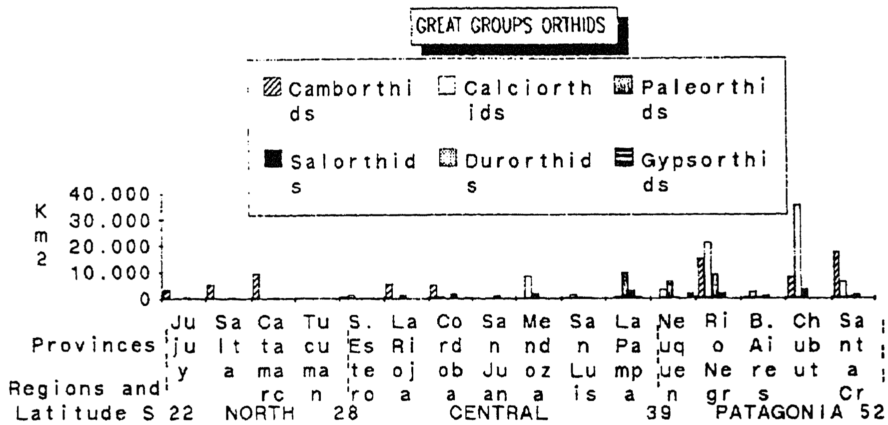
These appear in associations of different subgroups or are associated with Entisols. Alluvial or aeolian sediments from the foothills of high mountains are the parent materials for these soils. Lithic subgroups appear below 4,000 m, whereas, above this elevation (Puna and Prepuna), borollic subgroups are frequent. The typic and fluventic subgroups are found on alluvial fans and in the flat surroundings of drainage ways. The ustolic subgroup appears in higher, more moist places.

Salorthids

These appear associated with ustic Inceptisols in areas bordering salt deposits and fluvial surroundings, on plains, and in poorly drained depressions. The conductivity is higher than 60 mmhos\cm (Vargas Gil and Culot, 1980).



Graphic 2



Graphic 3

Camborthids

These appear as minor components in associations with Entisols and some Mollisols and also are found in association with other Aridisols. The parent materials are aeolian/ alluvial deposits in a plain landscape. Soft powdery lime appears under the diagnostic horizon.

Calciorthids

These are associated with Entisols and developed from aeolian sands recently deposited on silicified lime. These aeolian materials sometimes are mixed with other materials of a fluvial type (redeposited loess). These soils are found in distal areas of piedmont fans and relict areas of river beds and alluvial terraces.

Durorthids

These are associated with Entisols (Torriorthents) and developed in aeolian quartzitic sands in areas where the water table is close to the surface producing the saline characteristics and cemented horizons.

Gypsiorthids

These are primarily in the petrogypsic and typic subgroups. The parent materials are fluvial-lacustrian materials with a high content of calcium sulfate and soluble salts. They are common in areas near salt deposits.

Central Region

Haplargids

These are subordinated components of associations with Entisols and developed from alluvial sediments mixed with aeolian sediments in flat or concave-flat areas. In the particular case of the "barreales" (mudholes), the alluvial sediments brought coarser materials produced by erosion processes which occurred in higher areas.

Natrargids

These are associated with other Argids and/or Entisols and are developed in aeolian sediments in low areas with impeded drainage or in areas around salt deposits. These soils contain high values of soluble salts.

Salorthids

These are in the typic and aquollic subgroups and are associated with other Orthids. They have developed in fluvial-lacustrian materials overlaying sandy-loam and clayey sediments located in low areas of basaltic plains and in areas surrounding salt deposits and lagoons. Conductivity is higher than 35 mmhos/cm.

Paleorthids

These are associated with other Orthids and with Entisols. They are formed in loessal materials and are located in the higher portions of more or less defined hills with dendritic drainage patterns and usuallly contain lime layers. A petrocalcic horizon can be found at different depths.

Patagonic Region

A common characteristic of this region is the so - called "desert pavement," detritic cover due

TABLE 2: Geographic distribution of Argentina Aridisols.

1. SUBORDERS (%)

REGIONS	ARGIDS	ORTHIDS
Northern	11	10
Central	3	22
Patagonia	86	68

2. ARGIDS GREAT GROUPS (%).

REGIONS	PALEARGID	NATRARGIDS	HAPLARGIDS
Northern	35	1	3
Central	0	1	2
Patagonia	65	98	95

3. ORTHIDS GREAT GROUPS (%).

REGIONS	CAMBORTHIDS	CALCIORTHIDS	PALEORTHIDS
Northern	26	0	0
Central	16	14	39
Patagonia	58	86	61
	SALORTHIDS	DURORTHIDS	GYPSIORTHIDS
Northern	3	0	0
Central	58	100	12
Patagonia	39	0	88

TABLE 3

ABBREVIATED MORPHOLOGICAL, PHYSICAL AND CHEMICAL PROPERTIES OF SOME PEDONS OF THE GREAT GROUPS IN THE DIFFERENT REGIONS

Regions	Sub Groups	Horizons	Depth	Color	Granulometry			Organic	Nitrogen	$CaCO_3$	Extractable cations				CEC	Exchang	Conductivity	pH
					Sand	Silt	Clay	Carbon			Ca++	Mg++	Na++	K+		Sodium		
			cm	moist	%						meq/100 gr.					%	mmhos/cm	paste
CALCIORTHIDS																		
CENTRAL	TYPIC	A1	0-30	10YR 4/3	67,0	23,7	11,5	0,73	0,060	5,8	-	-	0,8	1,3	21,2	4,0	-	7,7
		AC	30-50	10YR 4/3	65,1	26,4	10,7	0,40	0,045	18,2	-	-	0,9	0,9	18,8	5,0	6,4	7,3
		Cca	50-100	10YR 5/3	60,9	26,0	11,7	0,18	-	10,0	-	-	2,2	0,8	21,2	10,0	7,5	7,4
PATAGONIA	TYPIC	A1	0-18	10YR 3/5	82,1	14,8	8,1	0,28	-	-	13,6	3,9	0,7	0,3	18,4	-	-	7,2
		AC	18-32	10YR 3/3	82,8	6,0	12,5	0,27	-	-	12,3	2,6	0,4	0,5	16,5	-	-	7,3
		2C1	32-45	10YR 3/2	70,1	21,2	17,2	0,66	0,076	0,5	25,6	5,6	2,5	0,4	32,2	7,0	-	7,6
		2C2ca	45-100	10YR 5/3	37,7	6,9	23,6	0,68	0,064	33,6	-	-	1,7	0,4	22,8	7,0	2,3	7,6
CAMBORTHIDS																		
NORTHERN	TYPIC	A1	0-20	10YR 6/2	80,8	11,4	8,8	0,32	0,040	-	3,03	0,76	0,24	0,38	3,36	7,14	0,25	7,0
		B2	20-50	7,5YR 5/4	77,2	12,6	9,2	0,09	0,028	-	-	-	0,25	0,38	4,54	5,51	0,28	8,1
		C	50-80	7,5YR 7/2	81,2	10,2	8,6	0,09	0,024	traces	-	-	0,33	0,29	3,42	9,65	0,26	8,3
CENTRAL	USTOLLIC	A1	0-22	10YR 3/2	5,0	68,0	28,8	2,67	0,299	-	36,7	2,4	0,2	1,9	33,8	-	-	7,8
		B2	22-60	10YR 3/2	10,0	61,8	30,9	1,70	0,178	-	29,3	2,4	0,1	1,5	30,1	-	-	6,8
		C	60-100	10YR 3/4	26,6	53,2	18,0	0,98	-	-	24,1	1,9	0,2	0,9	17,3	-	-	7,6
PATAGONIA	TYPIC	A1	0-27	10YR 4/3	62,7	22,9	14,3	0,34	0,048	-	13,8	13,18	0,56	1,38	16,55	3,40	-	7,5
		B2	27-40	7,5YR 4/4	65,7	19,1	15,1	0,29	0,038	-	23,0	3,06	0,85	0,92	22,20	3,83	-	7,6
		B3ca	40-50	7,5YR 5/4	54,1	12,1	1,2	0,14	0,025	3,9	-	-	0,63	0,62	17,18	3,68	-	8,0
		Cca	50-60	7,5YR 6/4	67,2	19,9	12,9	0,24	0,019	10,71	-	-	0,86	0,26	17,47	4,89	-	7,8
PALEORTHIDS																		
CENTRAL	TYPIC	A1	0-13	10YR 4/3	57,0	35,1	11,5	0,64	0,086	4,3	-	-	0,6	1,2	11,4	-	-	7,3
		2C	13-42	10YR 6/4	58,0	26,9	17,6	0,73	0,065	2,2	-	-	1,2	0,3	12,7	9,0	55,5	7,2
		tosca																
PATAGONIA	XEROLLIC	AC	0-18	10YR3/2	86,1	14,1	4,2	0,34	0,056	1,3	-	-	0,5	1,4	13,7	-	-	7,8
		C1	18-43	10YR 3/2	84,0	9,8	6,4	0,4	0,052	2,5	-	-	0,6	0,2	12,5	-	-	7,9
		C2	43-50	10YR 3/3	81,9	5,9	9,0	0,36	0,045	3,9	-	-	0,7	0,2	12,9	-	-	8,1
		tosca																
SALORTHIDS																		
NORTHERN	USTOLLIC	A1	0-20	10YR 8/2	63,0	28,0	9,0	0,42	0,038	1,1	-	-	0,7	0,13	4,6	15,2	66,0	7,9
		AC	20-40	10YR 3/2	54,0	33,0	13,0	1,37	0,085	11,2	-	-	2,2	1,59	16,0	13,8	67,0	7,8
		C	40-70	10YR 3/2	67,0	23,6	9,4	0,65	0,052	6,5	-	-	6,03	1,68	10,4	58,0	32,0	8,0
CENTRAL	TYPIC	I	0-25	7,5YR 4/4	14,3	86,5	15,1	0,51	0,033	1,3	-	-	5,8	1,3	15,0	39,0	83,3	8,0
		II	25-40	7,5YR 4/4	10,3	95,9	22,0	0,27	-	3,7	-	-	21,6	1,9	25,2	50,0	58,8	8,2
PATAGONIA	AQUOLLIC	I	0-20	5Y 6/3	7,4	18,8	10,8	0,06	0,005	1,9	-	-	7,2	0,6	15,2	47,0	8,3	9,2
		II	20-45	5Y 5/3	43,5	36,2	20,3	0,1	0,011	1,4	-	-	15,6	1,2	22,3	70,0	6,0	8,7
		III	45-100	5Y 5/4	51,4	35,6	12,9	0,28	0,028	19,6	-	-	25,0	1,4	29,9	84,0	32,2	9,4
DURORTHIDS																		
CENTRAL	TYPIC	AC	0-20	10YR 3/4	83,37	6,4	7,23	0,24	-	-	3,20	2,39	0,32	1,81	7,47	-	1,3	6,6
		C	20-40	10YR 3/4	86,14	6,97	7,59	0,14	-	traces	-	-	0,69	-	8,48	-	3,34	7,6
		2C2x	40-50	-	75,57	11,0	14,04	0,14	-	traces	-	-	3,02	-	16,64	-	9,24	8,0

to the selective removal and the thin discontinuous sandy accumulations.

Natrargids

These are associated with Entisols and with other Argids. They developed in aeolian sediments overlaying clayey sediments wich are more or less consolidated and can be found on various landscape positions: slopes, plains, tables, flats, and microdepressions.

Haplargids

These are associated with Entisols, Orthids, and xeric Mollisols. They have developed in aeolian sediments overlaying a clayey layer. They are found on plains, undulating tables, short steep slopes, and fluvial terraces.

Paleargids

They are associated with Entisols, other Argids, and xeric Mollisols. The parent materials are aeolian sediments of different thickness which overlay ancient clayey sediments. The concentration and hardness of calcium carbonate is variable. The position on the landscape is variable. A petrocalcic horizons occurs in some perfiles.

Calciorthids

They occur in association with other Orthids, Argids, and Entisols. They have developed in aeolian materials, on extended flat areas within the basaltic tables, on dissected smooth hills, and on marginal depressions of alluvial fans.

TABLE 3, CONT.

ABBREVIATED MORPHOLOGICAL, PHYSICAL AND CHEMICAL PROPERTIES OF SOME PEDONS OF THE GREAT GROUPS IN THE DIFFERENT REGIONS

Regions	Sub Groups	Horizons	Depth	Color	Granulometry			Organic	Nitrogen	Ca CO$_3$	Extractable cations				CEC	Exchang.	Conduc-	pH
					Sand	Silt	Clay	Carbon			Ca++	Mg++	Na++	K+		Sodium	tivity	
			cm	moist	%						meq/100 gr.					%	mmhos/cm	pasta
NATRARCIDS																		
NORTHERN	AQUIC	A1	0-18	10YR 2/2	82,2	14,00	3,8	0,46	0,058	1,0	-	-	0,38	1,12	8,1	4,6	1,15	8,06
		B2t	18-70	10YR 3/2	41,0	35,00	24,0	0,40	0,05	5,6	-	-	1,80	1,60	16,0	15,0	7,20	8,1
CENTRAL	TYPIC	Ap	0-18	10YR 3/3	58,8	43,00	13,8	1,03	0,045	-	11,8	3,4	1,30	1,70	16,1	8,00	-	7,5
		B2t	18-46	7,5YR 3/2	47,2	35,40	30,80	1,07	0,045	0,4	-	-	10,00	2,60	35,9	28,00	1,8	7,7
		B3ca	46-67	7,5YR 5/6	42,3	37,20	27,30	1,07	0,060	7,6	-	-	12,90	2,50	36,8	35,00	5,9	7,8
		Cca	67-80	-	40,8	18,80	26,30	1,05	-	20,4	-	-	12,90	1,9	30,2	43,00	10,9	7,8
PATAGONIA	TYPIC	A1	0-9	10YR 3/4	56,7	53,40	9,9	0,9	0,071	1,8	-	-	4,80	2,1	14,6	33,00	-	8,5
		2B2t	9-22	10YR 4/4	37,9	31,20	30,9	1,63	0,141	9,8	-	-	14,90	1,4	22,0	68,00	4,2	8,6
		2B31	22-34	7,5YR 3/4	65,9	1,50	32,6	0,28	0,029	17,9	-	-	10,90	0,3	26,0	42,00	22,5	8,4
		2B32	34-58	7,5YR 5/4	79,1	20,30	0,5	0,31	0,031	11,3	-	-	1,00	0,2	21,3	5,00	76,6	7,7
		2C2	58-100	7,5YR 6/3	83,3	5,70	10,9	0,12	0,010	11,4	-	-	2,90	0,2	18,5	16,00	13,1	8,0
PALEARGIDS																		
NORTHERN	TYPIC	A11	0-8	10YR 4/3	91,0	3,60	5,40	0,28	0,028	-	2,78	0,18	0,29	0,30	3,27	8,9	0,32	6,7
		A12	8-32	10YR 5/3	91,8	3,20	5,00	0,07	0,023	-	2,45	0,20	0,33	0,23	3,32	9,9	0,39	0,8
		B2t	32-55	5YR 4/3	54,6	4,00	41,40	0,42	0,057	-	12,86	3,95	1,16	0,94	19,80	5,9	0,45	6,8
PATAGONIA	XEROLLIC	A1	0-6	10YR 4/2	49,8	26,7	23,4	1,07	0,106	2,2	-	-	0,30	0,40	26,7	1,0	-	7,2
		2B2t	6-22	10YR 3/2	39,2	13,7	46,9	1,60	0,156	2,4	-	-	0,50	1,70	26,4	2,0	-	7,3
		3C	22-60	7,5YR 4/2	79,9	11,2	12,8	1,13	0,013	6,3	-	-	0,50	0,40	18,3	3,0	-	8,0
HAPLARGIDS																		
NORTHERN	BOROLLIC	A1	0-25	7,5YR 6/2	81,4	12,8	5,80	0,63	0,069	-	7,99	1,27	0,10	0,41	11,8	0,85	0,18	6,3
		B2t	25-50	5YR 4/4	35,4	41,2	23,4	0,28	0,041	-	15,40	3,70	0,23	0,39	21,2	1,08	0,27	6,3
CENTRAL	TYPIC	A1	0-12	10YR 6/2	45,8	51,1	21,1	1,08	0,120	0,3	-	-	0,70	2,90	23,1	-	-	7,6
		B21t	12-30	7,5YR 4/4	25,4	45,0	28,8	0,85	0,067	0,1	-	-	1,40	3,40	34,2	4,00	-	7,4
		B22ca	30-45	7,5YR 5/4	16,7	42,7	26,5	0,45	0,067	11,8	-	-	2,20	3,60	38,9	6,00	-	7,5
		B3ca	45-75	7,5YR 5/4	23,3	30,2	28,3	0,17	-	32,6	-	-	3,20	2,60	31,9	10,00	1,0	7,6
		Cca	75-100	7,5YR 6/4	54,0	27,2	21,1	0,08	-	10,7	-	-	6,50	2,30	35,3	10,00	2,4	7,7
PATAGONIA	TYPIC	A	0-11	10YR 5/3	83.1	7,2	9,7	0.92	0,069	-	6.4	1,60	0,3	0,8	15,3	2,0	-	7,7
		2B2t	11-33	7,5YR 3/2	54,5	15,2	30,6	1,69	0,143	2,7	-	-	0,3	0,4	33,1	-	-	7,7
		3Cca	33-100	10YR 8/2	67,1	19,6	13,3	1,00	0,079	48,2	-	-	0,8	0,1	18,1	4,0	-	8,9

Paleorthids

These are associated with other Orthids and with some Entisols. They have developed in aeolian sediments and clayey material on terraces, plains and tables.

Camborthids

They occur in association with other Aridisols and with Entisols. They have developed in aeolian/alluvial materials on table-shaped plains which are gently undulating, in extended depressions, and on gentle slopes.

Salorthids

These are associated with Argids and with some Entisols. They occur in depressions (salt deposits) on short gentle slopes and in poorly defined drainage ways consisting of depressions forming a line. The conductivity of these soils is 45 mmhos/cm.

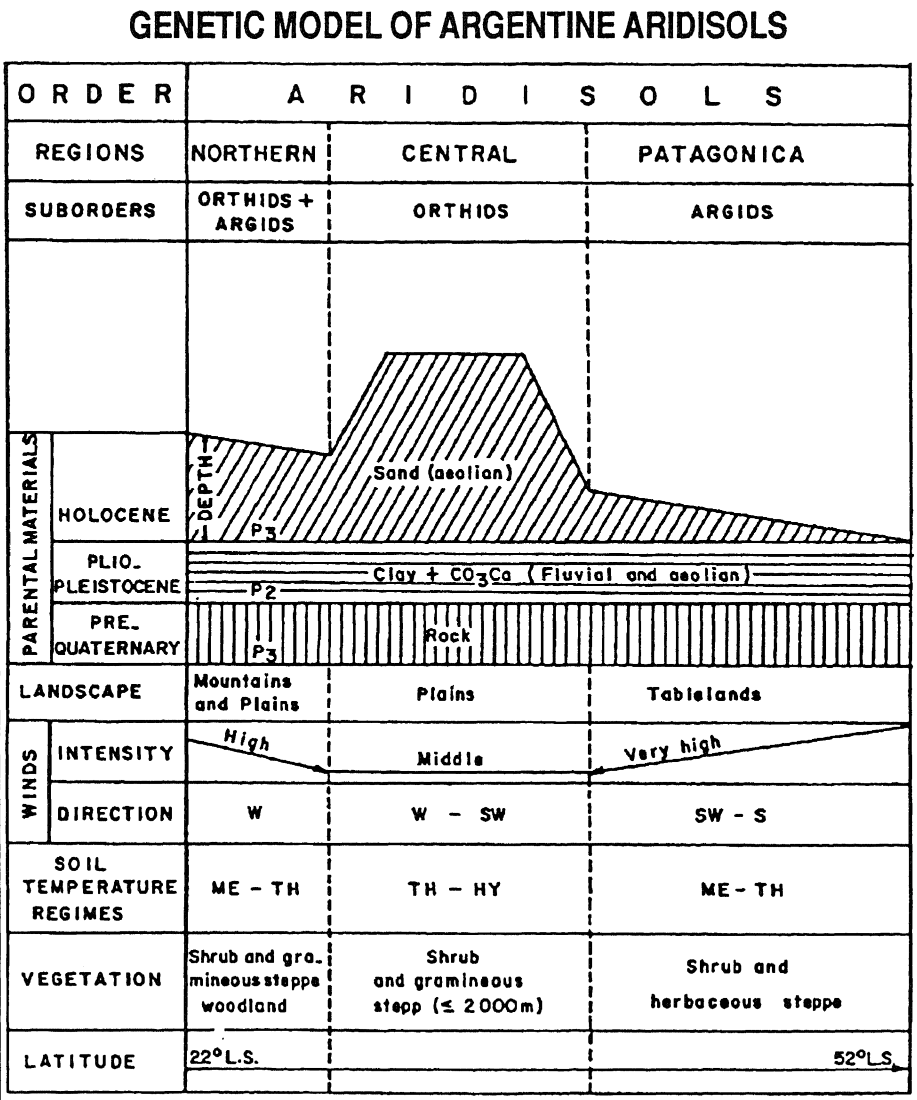
Gypsiorthids

They are associated with other Orthids (Salorthids). The parent materials are fluvial-lacustrian sediments with high contents of calcium carbonate. They are located in depressional areas, near salt deposits (Salazar and Godagnone, unpublished, 1984).

Genetic Considerations

The variety and distribution of the identified taxa as well as the peculiar geological and pedological characteristics existing in Argentina allow some possible conclusions about the genesis of these Aridisols. Graphic 4 presents, in a genetic model, these conclusions, which mainly refer to the origin and distribution of the Suborders (Argids and Orthids).

The starting point is to assume that the whole area with aridic SMR - and also most of the country - is covered by modern sediments (Pleistocene?, Holocene, and Recent) of aeolian origin. These sediments are friable, sandy and sandy loam textured material (P3 in the model) overlaying Pliopleistocene (P2) clayey materials. The thickness, distribution, nature, and grain size of these Holocene (P3) sediments are a result of the drifting by winds, blowing from the S-SW-W direction, which originated in the South



Graphic 4

Pacific anticyclone. These winds are extremely intensive in the Patagonic region, causing deflation and drifting of materials which later on are deposited in the central and northeastern regions that become accumulation areas as winds loose intensity. Thus, the Holocene and Recent sediments, which are very thin or practically nonexistent in the Patagonia, become very thick in the plains of the central region and is medium in the northern region according to the characteristic landscape of tables and valleys (Graphic 4).

The geological, physiographic, and pedological features indicate that the Holocene sediments - when thick enough - are the parent material of the Aridisols as a whole. On the contrary, these Holocene sediments are the parent material of surface horizons when interrupted by other underlaying materials such as the Pliopleistocene clays (P3) which would constitute the argillic horizon of the Aridisols. The clay and organic carbon contents of the Bt horizons in the Argids and its Great Groups show this lithologic discontinuity. Therefore, the origin of the argillic horizon of the Argids is due to environmental conditions which are different from the present ones and which were present during the formation of the P2 material. The present pedoclimatic conditions would have not influenced this argillic horizon but would have acted almost exclusively on the A horizon developed from the Holocene sediments (P3).

The presence of Argids on Orthids depends fundamentally on the thickness of P3 material, which is a consequence of the action, direction, and intensity of the winds (main sedimentological agent) - more or less constant since the end of the Andean orogeny - passing through different landscapes, climates and vegetation.

The differences at the Great Group level are due, in the case of the Argids, to the local characteristics of the P2 material (Pliopleistocene) which are a consequence of its own genesis. This is reflected in the variable content of calcium carbonate, gypsum, and sodium, grain size composition, and grade of cementation, the presence and/or quantity of which results in conditiona for the development of Paleargids, Natrargids, and Haplargids.

The taxonomic difference for Orthids can be explained by the aforementioned characteristics plus the presence of soluble salts in the Holocene sediments (P3), which are directly responsible for the presence or absence of Calciorthids, Gypsiorthids, Camborthids, Paleorthids, Durorthids and Salorthids. The very abundant Recent sediments resulting from the deposition of new ma-

terials, as well as the reworking of those of the Holocene, have generated different types of Entisols (Torriorthents, Torrifluvents, Fluvaquents and Torripsaments) which are no doubt the most common in the region with aridic SMR in Argentina.

References

Van Wambeke, A. and Scoppa, C.O. Las taxas climticas de los suelos Argentinos. Determinaci òn de las definiciones del Soil T axonomy mediante el modelo matem ätico de Newhall y su resoluci òn por computaciòn Fortran IV. RIA s3, v. XIII, N 1, pp. 7-39. 1976.

Cabrera, A. Regiones Fitogeogr àficas Argentinas. Fasc. 1, Tomo II Enciclopedia Argentina de Agricultura y Jardinerìa. 2da. Ediciòn. Ed. ACME SACI. Buenos Aires. 1976.

INTA. Atlas de Suelos de la Rep ùblica Argentina (in press).

Vargas Gil, R. and Culot, P. Los Suelos de la Puna. Actas IX Reunion Argentina de la Ciencia del Suelo, Tomo III, pp. 1.065-1.075. Pararà. 1980.

Salazar Lea Plaza, J.C. and Godagnone, R. Suelos de la Provincia de Rìo Negro (in press).

Classification of Aridic Soils, Past and Present: Proposal of a Diagnostic Desert Epipedon

A. Souirji[1]

Abstract

A diagnostic desert horizon, namely the torric epipedon, is proposed as an alternative to the use of the aridic moisture regime in the definition of high level taxa of aridic soils.

The strong variation in moisture and air content of the torric epipedon induces physical dispersion, which play a major role in current desert soil formation.

Eolian activity also plays an important role in desert soil formation, and the occurrence of eolian sand is proposed as a diagnostic criterion.

The low organic matter content of desert soils is due to low litter production, its dispersion by wind and surface wash, and low incorporation, rather than to high decomposition.

Introduction

Early authors on aridic soils considered that there is little chemical and biological weathering in deserts because of low precipitation and scarce vegetative cover. Hence, only coarse-textured, undifferentiated soils are to be found in these environments (Hilgard, 1906). These ideas stemmed directly from the concept of zonal soils at a time when little data were available on soils of the arid regions.

With the development of irrigation in arid regions, more knowledge was accumulated about aridic soils, and their diversity became evident. "All manner of soils are found in arid areas and that to refer to 'desert soils' as if they constituted a homogeneous group is both erroneous and misleading" (Jackson, 1957).

All major soil classification systems had high level taxa of aridic soils. Brown desert soils in the USSR (Tiurin, 1965) and Light colored soils of arid regions in the USA (Thorp and Smith, 1949) are two examples. Most of these classifications distinguished aridic taxa at the highest level merely by their geographical location in arid areas.

Some of the main characteristics of aridic soils, such as the presence of a vesicular crust, low organic matter content with dominance of fulvic acids, and surface accumulation of carbonates, were known to the Russian pedologists since the early decades of this century (Lobova, 1960).

However, the wealth of information gathered by the Russian pedologists did not lead to an effective classification of aridic soils because genetic factors of soil formation, soil processes, and geography were explicitly used in the definition of the taxa instead of the soil properties themselves.

The Australian classification before 1960 (Stephens, 1954), the Israeli soil classification before 1979 (Dan and Koyoumdjiski, 1963), the French classification (CPCS, 1967), and the American classifications before the publication of *Soil Taxonomy* (USDA, 1975) all were influenced strongly by the Russian soil classifications and had the same limitations.

The publication of *Soil Taxonomy* marked the start of a new era in soil classification. The main innovations that stimulated broad international acceptance were:

- The use of soil properties for classification.
- The use of precisely defined diagnostic criteria and horizons to define the taxa.
- The use of a rational nomenclature.
- The organization as a key.

However, the use of soil climate to define high level taxa has been criticized, and features associated with moisture regimes were proposed as a preferable alternative (Sombroek,1985).

The latest update of the FAO-UNESCO soil map of the world (FAO,1988) supressed all reference to soil climate in the definition of taxa. In lieu of that, a yermic phase was defined based on soil properties.

In this paper the use of the aridic moisture regime for classification purposes will be discussed first. Then desert soil forming processes and the associated soil properties will be reviewed, leading to the definition of a diagnostic epipedon.

[1]Soil Scientist, F.A.O., P.O. Box 5287, Sultanate of Oman.

The Aridic Moisture Regime

Definition of the Aridic Moisture Regime

In the aridic (torric) moisture regime, the moisture control section in most years is:

a. Dry in all parts more than half the time (cumulative) that the soil temperature at a depth of 50 cm is above 5°C; and

b. Never moist in some or all parts for as long as 90 consecutive days when the soil temperature at a depth of 50 cm is above 8°C (USDA-SCS, 1975).

"Soils that have an aridic or torric moisture regime are normally in arid climate." This statement extracted from *Soil Taxonomy* (USDA-SCS, 1975) emphasizes the close link which exists in most cases between the aridic soil moisture regime and the atmospheric climate. The question that arises then is whether or not the aridic moisture regime is a soil property. The analysis of the factors influencing soil moisture regimes in general may provide an answer to this question.

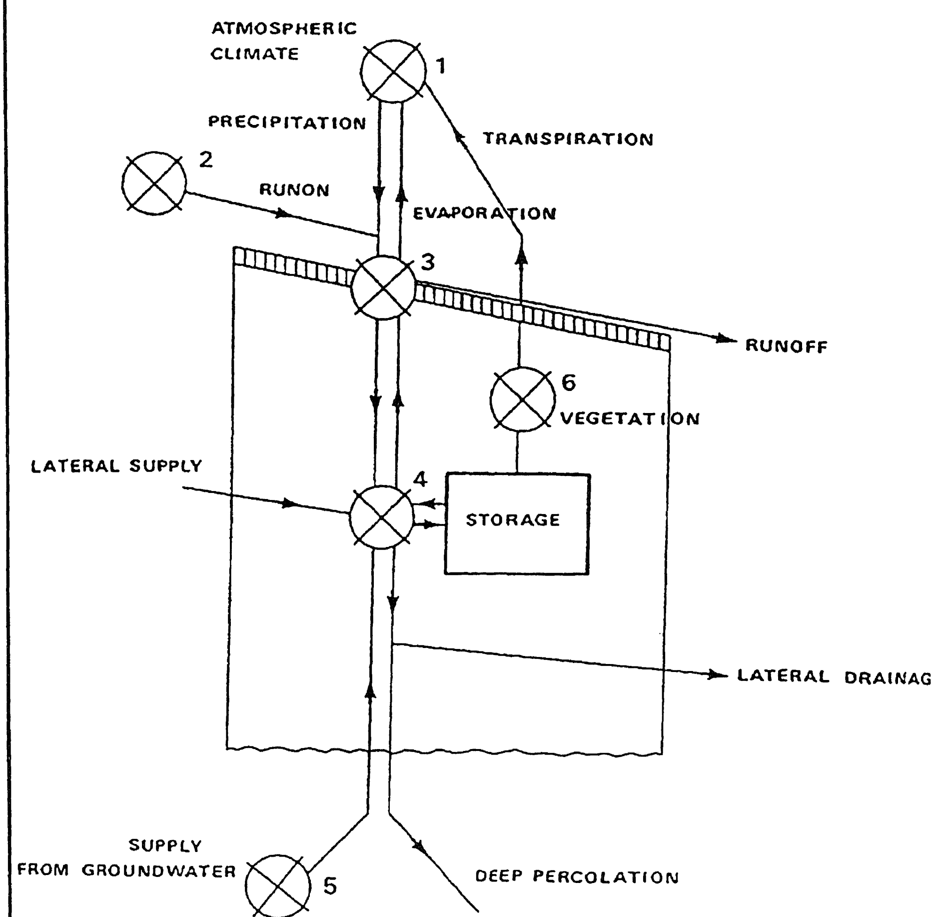


Figure 1: Factors influencing the soil moisture regime.

Factors Influencing the Soil Moisture Regimes

Figure 1. attempts to show the factors influencing the soil moisture regime as a system, by displaying the factors as a series of "valves".

Valve 1 represents the atmospheric climate which controls entirely the precipitation and strongly the evaporation and transpiration.

Valve 2 represents the geomorphological factors controlling runon (water runoff from adjacent soils).

Valve 3 represents the soil-atmosphere interface, i.e., the soil surface. It controls infiltration and evaporation. Slope and infiltration rate are the most relevant soil characteristics.

Valve 4 represents the soil characteristics that influence the transfer of water through it. Saturated and unsaturated hydraulic conductivity are the most important in this regard.

Valve 5 represents the hydrological and geomorphological factors controlling moisture supply from groundwater.

Valve 6 represents the vegetation factor which strongly controls transpiration.

Valve 7 represents the storage of water which is controlled by soil properties, e.g., water holding capacity.

It appears then that the moisture intake and storage by the soil is controlled by soil properties, whereas the moisture supply is essentially an independant variable. The equilibrium of this system and its pattern of change with time constitute the soil moisture regime.

It also appears from Figure 1 that, if valve 1 (atmospheric climate) is closed, the soil, whatever its properties and in the absence of groundwater supply, will remain dry.

The aridity of desert soils is directly caused by the aridity of the atmospheric climate without

the involvement of soil properties or pedogenic processes. However, in some areas bordering deserts and having a semi-arid climate, some strongly sloping or crusted soils may have an aridic moisture regime. Since these occurences are very limited in extent, it can be stated that the aridic moisture regime is a property of the climate rather than of the soil.

The Determination of Soil Moisture Regimes

Although moisture regimes can be measured *in situ*, the procedure is costly, time consuming, and unsuitable for soil surveys, since at least one decade of measurement would be necessary to obtain statistically significant results.

Therefore, mathematical models were developed to compute moisture regimes from atmospheric climatic data. The model of Newhall (1980) is widely used. The only soil property taken into account in this model is the water holding capacity. The soil is considered as an inert reservoir; therefore the variations of the moisture storage are entirely determined by the atmospheric climate.

It is likely that, in the future, better models will be elaborated, but they always will depend on climatic data, which are scarce in most arid areas.

The Use of the Moisture Regime for Soil Classification

Soils are the objects to be classified; therefore only soil properties are to be used as diagnostic criteria.

Because of the important bearing of the moisture regime on soil genesis, moisture regime can be used to decide which soils belong together. The next step is to find out associated morphometric criteria to define the grouping. More refined classes of moisture regimes could be used at the family level to enhance the agronomic significance of the lower taxa (Sombroek, 1985). Indeed, the current classes of soil moisture regime are too broadly defined to allow useful agronomic interpretations within the same climatic zone.

Desert Soil Forming Processes and Associated Properties

Strong moisture deficit, extreme temperatures, and high wind activity are the main factors controlling soil formation in deserts. These factors, alone or in combination, are reflected in desert soils' composition, morphology, and behavior.

Organic Matter In Desert Soils

In a virgin soil, the organic carbon content reflects the equilibrium between (mostly) vegetal biomass production and decomposition.

Vegetal biomass production depends on the availability of moisture and heat for a duration long enough for the vegetation to accomplish its biological cycle.

In some deserts such favourable conditions occur in most years, whereas in others they occur rarely and only locally. The dynamics of organic matter would therefore be quite different in deserts with a regular growing season and those without.

Deserts With a Regular Growing Season

These deserts are, rather, semi-deserts and are encountered in areas transitional to semi-arid. Some of the North-American deserts, such as those of Wyoming or Idaho, belong to this category. Precipitation is rather high, often more than 200 mm, and rainfall occurs in spring and early summer when soil temperature is optimal. As a result, a dense vegetation of shrubs, forbs, grasses, and mosses develops.

A sod can form and litter accumulates in the topsoil, protected from wind deflation by the dense bush cover. The soil fauna can also develop and contribute to the decomposition and incorporation of organic matter in the soil.

In these semi-deserts, high amounts of organic carbon, often more than 2-3 percent, occur in the A horizons.

Deserts Without a Regular Growing Season

These are the true deserts. They are either hot or cold but the amount of precipitation, generally well below 200 mm, and the rainfall pattern are such that moisture supply to the vegetation is both limited and erratic.

In these conditions, a sparse woody vegetation of shrubs tends to be dominant, to the detriment of grass, and a sod cannot form. Due to the scarcity and type of vegetative cover, the rate of litter production is very low and removal by wind and surface wash strongly reduce its *in-situ* accumulation.

The initial decomposition of litter into smaller particles is normally done by insects, collembola, and other small invertebrates before further decomposition by microorganisms occurs.

These primary decomposers are very scarce in true deserts.

The final decomposition of organic matter is governed by microbial activity which itself depends on the available moisture, aeration, and temperature. Since the soils are dry most of the time and the temperature is often either very high or very low, desert environments are not, except for aeration, favorable to high decomposition rates. Birch (1958) showed that, when a dry soil is moistened, a flush of decomposition occurs. Hence, the alternation of drying and moistening would favor rapid decomposition of organic matter. However, in true deserts rainy episodes are usually rare and brief, and therefore the global rate of decomposition is rather low.

Earthworms, which usually contribute significantly to the incorporation of organic matter into stable organomineral compounds, are totally absent from most desert soils.

It seems, then, that the low organic matter content of true desert soils is primarily due to low litter production, dispersion, and low incorporation, rather than to high decomposition.

In true deserts the organic carbon content of the A horizons is usually well below 0.6 percent, and if the texture of the A horizon is loamy sand or coarser the organic carbon content is 0.2 percent or less (Souirji, 1987b; FAO, 1988). However, in the driest deserts, such as those of the Arabian Peninsula or the Sahara, the organic carbon content of the A horizons is often less than 0.1 percent no matter what the texture is (Dutil, 1971; Souirji, 1990 in prep.). These values are in good agreement with previously published data (Lobova, 1960; USDA, 1975; Kovda et al., 1979).

Organic carbon content increases with moisture availability owing to poor drainage, irrigation, or both. It then can be higher than the above-mentioned values, but if so salinity increases beyond 4 ds/m (saturated paste extract) within 125 cm below the soil surface (Souirji, 1987 b; FAO, 1988).

Exceptions are encountered, however, in cases of soils having a good internal drainage flooded with low salinity waters. In virgin desert soils, this situation was only found to occur in some soils of flooded depressions (Souirji, 1990 in prep.).

It is noteworthy that organic matter contents can be higher in the subsoil than in the topsoil and that irregular distribution in the profile is common.

The organic matter of typical desert soils contains more fulvic than humic acids and more waxes and resins than other soils (Lobova, 1960; Kovda et al., 1979). Iron, rather than calcium, seems to be the bonding agent between the organo-mineral compounds (Lobova, 1960).

Physico-Chemical Behavior of the Air-Liquid-Solid Phase System

In temperate areas and the humid tropics, the relative humidity of the air is usually high, around 70 percent or more. Lower relative humidity does occur but only for short durations.

Except in some coastal areas, air relative humidity is low in deserts. For example, in the arid part of Saudi Arabia, mean monthly relative humidity is below 45 percent most of the year and below 25 percent for several consecutive months (Al-Zeidi et al., 1988).

The pF of a soil in equilibrium with such relative humidity (less than 48 percent) is above 6. Therefore all the pores having a diameter of 15 microns or more are filled with air (Tessier, 1984).

A very high air content and a low water content for extensive durations are very important characteristics of aridic soils. They are responsible for the occurence of physical dispersion and of a special type of chemical weathering.

Physical Dispersion

Structural crusts occur in soils of various climatic areas but are much more widespread in arid and semi-arid regions. "They are normally more compact, harder and more brittle than the soil just beneath them" (Evans and Buol, 1968).

Structural crusts can form under the influence of externally applied mechanical pressures (e.g., rain drops impact) or through the rearrangement of the soil fabric following slaking upon wetting.

It has been shown experimentally that the slaking of the soil upon wetting is due to either compression of air in the pores or to rapid swelling or both (Quirk, 1950; Tessier, 1984; Le Bissonais, 1989) .

Compression of Air

The above-mentioned authors and others have shown that, the lower the initial soil water content, the stronger its tendancy to slake upon wetting. By wetting the soil alternatively under atmospheric pressure and under vacuum, they have found that the compression of the air filling

the pores by the advancing moisture front causes the collapse of the aggregates. The intensity of slaking by air compression depends on the soil air content, which is determined by the climate and the pore-size distribution.

In temperate climates, structural crusts occur mostly in silty soils, because these have larger pores that can dry out and become filled with air even during moderately dry climatic episodes, e.g., European summers.

The surface soil in arid zones is usually extremely dry (filled with air) for extensive periods and therefore presents ideal conditions for slaking by air compression. Hence, structural crusts occur in a much wider range of textures than in humid areas (Souirji, 1990 in prep.).

Structural crusts of arid regions are generally vesicular (Jackson, 1957; Springer, 1958; Lobova, 1960; Miller, 1971; Kovda et al., 1979; Nettleton and Peterson, 1983; Souirji, 1987b, 1990 in prep.). This peculiar feature is due to the fact that arid soils have a much higher air content and that the uppermost part of the crust dries out immediately after the rain stops, hence trapping the air which is bubbling up in the dispersed soil.

Swelling

Clay minerals are known to have an internal deficit of positive electrical charges caused by isomorphous substitution of Si^{4+} and Al^{3+} ions by other cations of lower valence. The electrical deficit is compensated for by the adsorption of counter ions of an equal opposite charge on the surface of the mineral.

In presence of water, the adsorbed cations tend to diffuse away, but the electric attraction of the negatively charged mineral's surface forces them to remain in an adjacent layer of the soil solution. A diffuse double layer (D.D.L.) is formed which differs from the outer soil solution in that it contains a large excess of cations over anions.

In case water is removed from the soil to the point that the D.D.L. shrinks below its potential thickness, we are in the presence of what Bolt (1976) has called a truncated diffuse double layer (T.D.D.L.). The greater the difference between the potential (maximum) extent and the actual extent of the D.D.L., the stronger the tendency of the system to expand.

Translated in terms of physical behavior, the system is similar to an osmometer, and it will develop, upon wetting, a very strong swelling pressure of the order of several tens of bar. This swelling pressure is higher if (1) the adsorbed counter ions are monovalent, (2) the soil solution has a low electrolyte concentration, e.g., rainwater, and (3) higher specific area (e.g., 2/1) clay minerals are present (Bolt, 1976).

In the arid regions, conditions are favorable for the occurence of T.D.D.L because the soils are usually very dry and sodium and potassium have a larger share of the base saturation than in humid areas.

Consequences of Physical Dispersion

Strong physical dispersion has the following consequences:

- weak structure (often platy) in the topsoil and crusting
- higher mobility of the clay fraction which can be leached (Souirji, 1990 in prep.).

Indeed, strong dispersion of the clay fraction, if the topsoil is permeable, may cause the formation of shallow argillic horizons. Also, since the peds are dispersed, clay coatings cannot form.

Chemical Weathering

In a soil that becomes very dry, the residual water has properties very different from an ordinary soil solution.

The hydrogen atoms of the soil water molecules become about one million times more mobile, hence increasing their chemical activity, which may become equivalent to an H^+ concentration of about 0.5 N (Chaussidon and Pedro, 1979).

This strong variation of the pH of the soil solution, which also is accompanied by a variation in cations and anions concentration, may induce selective exchange with cations which are in isomorphic substitution (e.g., adsorbed Mg exchanging Fe).

The elevation of the pH of the soil solution by two or more units upon wetting also may enhance weathering of silica. The occurence of amorphous silica films on quartz particles of desert eolian sands is well documented (Le Ribault, 1977). If glass or opal (e.g., from roots decay) are present, one can speculate that, under an arid climate, marginal to ustic or xeric, weathering and leaching of silica may become strong enough to form duripans.

Varnish on pebbles and stones of desert pavements also is probably formed by weathering caused by residual water from evaporation of moisture films deposited by the dew (Souirji, 1990 in prep.).

Aeolian Additions and Turbations

The scarcity of vegetation and the occurence of strong thermic gradients cause the formation of violent and frequent winds that are characteristic of deserts.

Because of the existence of large areas of barren rocks undergoing physical weathering, desert winds carry large amounts of sand, which is redeposited as dunes or more commonly as small hummocks around vegetation or as thin sand 'veils' on the soil surface.

Playas and wadi beds can contribute substantial amounts of airborne clay and silt, which are redeposited in the neighboring areas or hundreds of kilometers away. Aeolian additions of clay, silt, sand, and salts to the soils are well documented (Lobova, 1960; Yaalon, 1973; Nettleton and Peterson, 1983; Souirji, 1987a and b).

Aeolian sands have a peculiar morphology (Cailleux et Tricart, 1959; Le Ribault, 1977). The individual quartz and feldspar particles are rounded (not necessarily spherical) and have a mat (dull) surface. Roundness is more frequent in particles larger than 0.250 mm. Souirji (1987a and b) proposed to use the occurrence of such aeolian sand as a diagnostic criterion for aridic taxa.

Ventifacts, varying from perfect dreikanters to glossy abraded pebbles, commonly are found in desert pavements. The degree of wind shaping depends on the age of the pavement and its lithology.

Recently deposited sandy alluvium often is reworked by wind and shows characteristic 'eoturbations' in the form of cross stratifications (Souirji, 1987b).

Salt Accumulation

Because of limited leaching, the soils of arid regions often contain accumulations of carbonates, sulfates, chlorides, and nitrates. These accumulations can occur only if there is a source of the above-mentioned salts. The sources may be allochthonous (e.g., airborne dust) or autochthonous (e.g., parent materials or shallow water table).

Differences in solubility product make the carbonates accumulate at shallow depth, whereas sulfates, chlorides, and nitrates accumulate at increasing depth. However, this is only valid for simultaneous deposition, and different salts can be found in the same horizon if they were deposited under different moisture regimes.

Many desert soils contain shallow lime accumulations separate from deeper relict calcic horizons (Figure 2). These 'duplex' profiles are quite common (Lobova, 1960; Souirji, 1990 in prep.).

Gypsum accumulations, when they occur, often underlie calcic horizons.

Acicular Clay Minerals

The presence of acicular clay minerals, mostly palygorskite, in desert and semi-desert areas was reported by many authors (Millot et. al., 1969; Eswaran and Barzanji, 1974; Singer and Norrish, 1974; Aba-Husayn and Sayegh, 1977).

Some authors attributed their origin to authigenic formation, whereas others favored inheritance from the parent material or windborne additions. Conrad (1969) and Dutil (1971), who studied the soils of the Algerian Sahara, found a clear association between palygorskite occurence and sebkha (playa) or fluviolacustrine deposits.

Zelazny and Calhoun (1977) identified prerequisites for the occurence of palygorskite and sepiolite as a relatively closed soil system with minimal leaching with little or no influx of Al and low H_3O but high Si and Mg concentrations. These conditions are fulfilled in many playas and temporary lakes of the arid zone.

Indeed, deserts, which generally have an endoreic (not draining to the sea) hydrographic system or lack one, are extremely favorable to the formation of playas and lakes, especially when moister climatic periods follow drier ones. Such a climatic alternation is known to have occured in most deserts; therefore, palygorskite-rich sediments could be quite common in these environments.

These sediments can become palygorskite-rich soils and/or a source of palygorskite- (sepiolite-) rich airborne dust to surrounding areas.

However, the occurrence of weathering in surface horizons of desert soils, bringing in the soil solution soluble silica and magnesium, may justify the existence of pedogenic palygorskite.

Whatever their origin, acicular clays when present in soils are a sign of very limited leaching, i.e., of aridity, unless they are protected inside hard, impervious structures such as nodules or petro-calcic and petrogypsic horizons. Therefore, the presence of palygorskite has been proposed as diagnostic criterion to be used in conjunction with other criteria to define aridic taxa (Souirji, 1987a; FAO, 1988).

Color of Desert Epipedons

Desert soils generally have light-colored surface horizons. This is probably due to low organic matter content and its type, the presence of iron oxides, and, often, a high lime content.

A review by Souirji (1990, in prep.) has shown that desert A horizons generally have Munsell color value 3 or more when moist and 4.5 when dry, and a chroma of 2 or more when moist.

Proposed Desert Epipedon

General Considerations

The shallow penetration of moisture in desert soils restricts current soil development to the upper horizons. Pedogenic development in deeper soil horizons of well drained desert soils is generally a relict of past moister climates. Epipedons are therefore the best markers of current soil fomation in deserts.

A definition for a diagnostic desert epipedon is given in the next paragraph. This definition is designed the same way *Soil Taxonomy* (USDA, SCS, 1975) defines the mollic epipedon, in order to show similarities in the rationales.

Definition of the Torric Epipedon

Abstraction of properties common to 'mature' automorphic soils of the drier arid zones focuses immediate attention on the horizons at or near the surface rather than deeper ones. Virtually all these soils exhibit a relatively thin, light-colored, humus-poor surface horizon or horizons in which divalent cations are dominant on the exchange complex and the grade of structure, often platy parting to subangular blocky, is weak to moderate. This kind of horizon may be named the torric epipedon.

From a genetic point of view, the properties of the torric epipedon are thought to develop under the influence of extreme dryness, extreme temperatures, and high wind activity. The formation and accumulation of stable organo-mineral compounds are limited and a high percentage of the organic matter is fulvic acid, waxes, and resins.

The soils having a torric epipedon have a widely varying mineralogical composition of the clay fraction but palygorskite, and less so sepiolite, is frequently encountered.

These soils are too dry for the cultivation of crops unless irrigated. Under cultivation, they have an adverse physical behavior, slake upon wetting, and are highly subject to wind erosion.

Although the torric epipedon is a surface horizon that can be truncated by erosion, its many important accessory properties suggest its use as a diagnostic horizon at a high categorical level.

The torric epipedon is defined in terms of its morphology rather than its genesis. It consists of mineral soil material. It is a surface A horizon or horizons unless it underlies a deposit of largely unaltered new material less than 50 cm. thick. It does not qualify as an anthropic, histic, mollic, plaggen, or umbric epipedons and has the following properties:

1. The organic carbon content is less than 0.6 percent if the texture is loamy very fine sand or finer and less than 0.2 percent if coarser (weighted average). If the soil has, somewhere within 125 cm below the surface, an electrical conductivity of the saturated paste extract of 4 ds/m or more, the conditions on organic carbon are waived.
2. Both broken and crushed samples have Munsell color value 3 or more when moist and 4.5 or more when dry, a chroma of 2 or more when moist.
3. Soil structure is weak or moderate and the consistence is soft when dry, although individual peds may be hard.
4. After a rain or an irrigation, a structural crust forms at the surface of the aridic epipedon. It generally has vesicular pores and a sandy loam or finer texture.
5. Base saturation is at least 75% by NH4 OAc method.
6. If the torric epipedon is calcareous, then it contains evidence of lime segregation in the form of coatings below gravels and coarse sand, pseudomycelia, or soft accumulations.
7. If there is a source of sand in the landscape, the sand fraction in some subhorizon or in inblown material filling the cracks contains a noticeable proportion of rounded or subangular sand particles showing a mat (dull) surface. These particles make 10 percent or more of the medium and coarser quartz sand fraction.
8. If a surface accumulation of coarse fragments (pavement) is present, some of them are shaped by the wind or show the presence of iron and manganese oxides accumulation (desert varnish) on their exposed surfaces or gypsum/calcium carbonate/sodium chloride below.

10. The lower limit of the torric epipedon is the upper limit of underlying B or C horizons. If the soil is cultivated and the Ap lies directly on a B or C horizon, the lower limit of the plow layer is also the lower limit of the torric epipedon.

Description of a Typic Torric Epipedon

A fully developed torric epipedon (Figure 2) has the following succession of features:

- A pavement composed of one or two layers of pebbles or stones, often wind-shaped (1) and/or varnished. Most of the coarse fragments are embedded in the underlying vesicular crust, but some of them are 'floating' in loose sand. In old pavements the lower part of the coarse fragments is often weathered and has an orange tinge or accumulation of gypsum (2).
- A sandy layer 0.5 to several cm thick (3). The upper part is often coarse sand and the grain size becomes gradually finer below. This sandy layer is sometimes discontinuous or absent.
- A thin layer 1 to 3 mm thick, generally massive, composed of clay, silt, and fine sand infiltrated from above (4). It is difficult to distinguish from the under-lying vesicular crust except above the cracks.
- A vesicular crust 0.3 to a few cm thick, having the same textural composition as the layer above (5). The vesicles are generally 1 to 3 mm in size. This crust is frequently thinly stratified and its transition with the underlying horizon is generally, but not always, abrupt.
- Wedged-shaped cracks (6), often filled with inblown sand, divide the above-mentioned crust and the underlying horizon(s) into polygonal prisms. When the sand 'veil' is removed carefully, the soil surface appears patterned into polygons by thin cracks (7).
- A weak to moderate platy (8), often parting to fine subangular blocks, horizon. Its thickness is generally less than 10 cm but can sometimes be as thick as 25 cm. When it is calcareous, lime segregation occurs in the form of pseudo-mycellia, coatings below pebbles, or soft accumulations, which generally increases in the lower part of the epipedon (9), which is often an Ak horizon.
- The lower horizons of old desert soils often contains a second, generally better developed, calcic horizon (10).

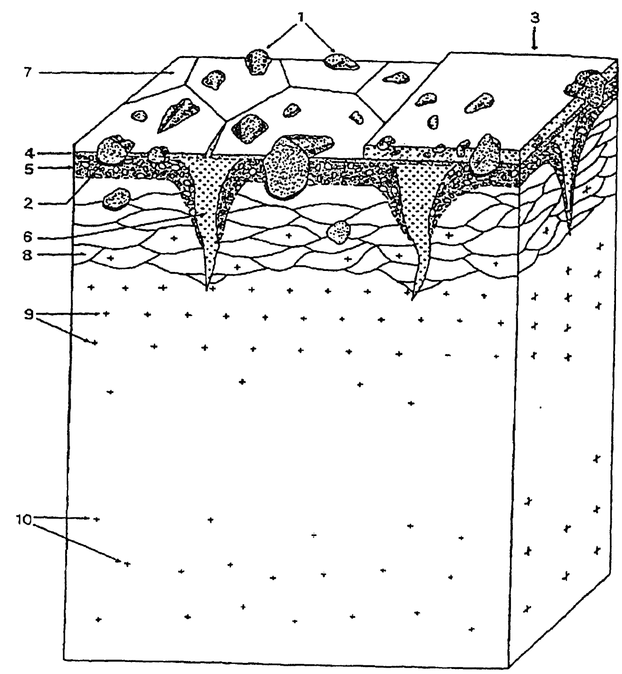


Figure 2. Morphology of a typical desert soil. (For legend, see text.)

Identification of the Torric Epipedon

The USDA (1975) defines recently deposited soil material (C horizon) as being thinly stratified. This definition is not adequate for sandy eolian materials because mass flow destroys the stratifications in very little time. Therefore, the absence of thin stratifications alone is not enough to indicate the presence of an A horizon.

The mere presence of roots in a sandy surface horizon is also not sufficient to indicate enough soil development to form an A horizon, because roots can form anywhere moisture, nutrients, and air are available.

The accumulation of organic matter in desert surface horizons being generally very limited or even less than in subsurface horizons, it cannot be used as criterion to define an A horizon.

In deserts, as soon as a surface soil horizon becomes stable, dust deposition and physical dispersion induce the formation of a structural crust. This structural crust controls all subsequent soil developement and gradually the other properties of the torric epipedon appear (Souirji, 1990 in prep.). Actually the only type of A horizon that can form in desert environments is the torric epipedon, which can be considered the desert ochric epipedon.

In the definition of the torric epipedon, reference is made to "rounded or subangular sand

particles showing a mat (dull) surface.....make 10 percent or more of the medium and coarser non-carbonate sand fraction." Since this determination is not a routine procedure in soil science, it is useful to give some details.

A composite sample of the A horizons is washed with water on a 250 microns mesh sieve to keep only the medium and coarser sand fraction. The sand is then put in a beaker and boiled with enough HCl acid to eliminate carbonates and iron oxides. The composite sample weight has to be adjusted in order to obtain about 100 sand grains.

The clean sand grains are then mixed and a subsample of about 30 grains is examined with a strong hand lense (20X) or a binocular microscope. In Saudi Arabia at least 80 percent of the sand grains have a sub-angular or rounded and mat surface (Souirji, 1987, 1990 in prep.). If eolian sand is encountered in the landscape, the presence of mat sand can be assumed.

Conclusions

Soils having a torric epipedon generally belong to the current order of aridisols (U.S.D.A., 1975) and have subsurface diagnostic horizons. However, entisols of stable geomorphic surfaces, mostly torriorthents and torripsamments, do have a torric epipedon. The author suggests keeping these soils in the order of entisols and defining special great groups for them. Indeed, the presence of a torric epipedon has an important genetic and management significance and therefore should be reflected in soil classification and subsequently in soil maps.

The use of the torric epipedon allows us to substitute field observation and simple laboratory determinations for uneasy soil climate measurements. It has also the advantage of drawing attention to the properties of the topsoil which too often are neglected in the profile descriptions. Structural crusts are a good example of features that are often overlooked, although they reduce infiltration and can hamper or prevent seedling emergence.

References

Aba-Husayn, M. and Sayegh,A., 1977. Mineralogy of Al Hasa Desert Soils. Clay Minerals, 25: 138-147.

Al-Zeidi et al, 1988. Guide for crop irrigation requirement in the Kingdom of Saudi Arabia. FAO - Ministry Of Agriculture And Water, Riyadh.

Birch, H.F., 1958. The effect of soil drying on humus decomposition and nitrogen availability. Plant and Soil, 19, 9-21.

Bolt, G.H. and Bruggenwert, M.G.M. (editors), 1976. Soil Chemistry A Basic Elements. Elsevier.

Cailleux, A. et Tricart, J., 1959. Initiation a l'etude des sables et des galets, 3 tomes. Centre Docum. Univers. Paris.

Chaussidon, J and Pedro, G. 1979. Role de l'etat hydrique du systeme poreux sur l'evolution du milieu. Realite de l'alteration dans les systemes a faible teneur en eau. Sciences Du Sol, numero double 2 et 3.

Conrad, G., 1969. L'Evolution Continentale Post-Hercynienne Du Sahara Algerien. Ph.D. thesis. Editions Du Centre National De La Recherche Scientifique, Paris, France.

C.P.C.S., 1967. Classification Des Sols. Laboratoire de Geologie-Pedologie E.N.S.A. Grignon. Multi. France.

Dan, J. and Koyumdjisky, H., 1963. The soils of Israel and their distribution. J. Soil. Sci., 14.

Dutil, P. 1971. Contribution A L'Etude Des Sols Et Des Paleosols Du Sahara. Ph.D. thesis. Faculte Des Sciences de L'Universite de Strasbourg, France.

Eswaran, H. and Barzanji, A.F. 1974. Evidence for the Neoformation of Attapulgite in Some Soils of Iraq. Trans. 10th Int. Cong. Soil Sci., Moscow, 7.

Evans,D,D and Buol,S,W, 1968. Micromorphological study of soil crusts. S.S.S.Am.Proc. Vol. 32 no 1.

FAO., 1974-1981. Soil map of the world, vol. I-X, Unesco, Paris.

FAO, UNESCO, ISRIC, 1988. Soil map of the world, Revised Legend. World Soil Resources Report 60. FAO, Rome.

Hilgard, E.W., 1906. Soils. McMillan, New York.

Jackson, E.A., 1957. Soil features in arid regions with particular reference to Australia. J. Austr. Inst. of Agric. Science. 196-209.

Kovda, V.A., 1979. Soils processes in arid lands in Arid-Land Ecosystems, vol. 1. Goodall, D.W. and Perry, R.A. (editors). International biological Programme, 16. Cambridge University Press.

Le Bissonais, Y, 1989. Analyse des processus de microfissuration des agregats a l'humectation. Sciences Du Sol, Vol. 27 no 2.

Le Ribault, L., 1977. L'exoscopie des quartz. Masson, Paris.

Lobova, E.V., 1960. Soils of the Desert Zone of the U.S.S.R. Issued in translation from Russian by Israel Program for Scientific Translations, Jerusalem, 1967.

Miller, D.E., 1971. Formation of Vesicular structure in soil. S.S.S. Am. Proc. Vol. 35, 1974. 635-637.

Millot, G. et al., 1969. Neoformation de l'attapulgite dans les sols a carapaces calcaires de la Basse Moulouya (Maroc Oriental). C.R. Acad. Sc. Paris, t. 268.

Nettleton, W.D. and Peterson, F.F., 1983. Aridisols In: Wilding, L.P., Smeck, N.E. and Hall, G.F, (editors). Pedogenesis and soil taxonomy, II. The soil orders. Elsevier.

Newhall, F., 1980. Calculation of soil moisture regime from climatic records. SMSS roneo. SCS, USDA, Washington.

Pye, K., 1987. Aeolian Dust And Dust Deposits. Academic Press.

Quirk,J,P, 1950. The water stability of soil aggregates in relation to the water content at the time of wetting. C.S.I.R.O. Division of Soils, Report 12/50.

Singer, A. and Norrish, K., 1974. Pedogenic Palygorskite Occurences in Australia. American Mineralogist, vol. 59.

Sombroek, W.G., 1985. A quest for an alternative to the use of soil moisture regimes at high categorical levels in Soil Taxonomy. In Proc. of the 5th Intern. Class. Workshop, Sudan, 1982, Part I. Soil Survey Administration, Sudan.

Springer, M.E., 1958. Desert Pavement and vesicular layer of some soils of the Lahontan basin, Nevada. S.S.S. Am. Proc. 22, 63-66.

Souirji, A., 1987a. Classification of Desert Soils. Land and Water Newsletter, 2, 1987, FAO, Rome.

Souirji, A., 1987b. Classification of Desert Soils. Ministry of Agriculture and Water, Project FAO. UTFN/SAU/ 015, Riyadh, Saudi Arabia.

Souirji, A., 1990 in prep. Contribution To The Study Of Desert Soils, The Torric Epipedon. Ph.D. thesis. Faculte Des Sciences Agronomiques de L'Etat a Gembloux, Belgium.

Stephens, C.G., 1954. The Classification Of Australian Soils. Proc. 5th Intern. Cong. Soil Sci. (Leopoldville), 4.

Tessier,D, 1984. Etude experimentale de l'organisation des materiaux argileux. Doctoral thesis. I.N.R.A., Paris.

Thorp, J. and Smith, G.D., 1949. Higher categories of soil classification. Soil Science, 67: 117-126.

Tiurin, V., 1965. The System of Soil Classification in the USSR. Main Stages in the Development of the Soil Classification System in the USSR. Pedologie, spec. 3.

USDA. Soil Survey Staff, 1975. Soil Taxonomy. U.S. Government Printing Office, Washington.

Yaalon, D.H. and Ganor, E. 1973. The influence of dust on soils during the quaternary. Soil Sci., 116.

Zelazny, L.W. and Calhoun, F.G., 1977. Palygorskite, Sepiolite, Talc, Pyrophyllite, and Zeolites. In "Minerals in Soil Environments" J.B. Dixon et al. editors, S.S.S. Am.

Report on Site Specific Soil-Climate-Vegetation Relationships: Cold Vertisols-Aridisol ISCOM Tour (Montana - Idaho - Wyoming)

G.J. Staidl and S.G. Leonard[1]

Abstract

Rangeland vegetation community characteristics are highly varied due to the complex interaction between soil, ambient climate, and other environmental factors with specific vegetation requirements and community relationships. This study was conducted to investigate and develop soil-vegetation relationships for the rangeland soil sites of the Idaho, Montana, and Wyoming part of the Cool Vertisol-Aridisol ISCOM Tour. The results were incorporated in the tour guide for evaluation and use in the *Soil Taxonomy* revision process. Documentation included soil pedon descriptions, soil characterization data, vegetation characteristics, climate data, and supporting literature.

An on-site examination in the spring, 1989, emphasized the affect of soil physical and chemical properties in relation to climate on plants and community characteristics. Each site reflected its own unique soil-climate-vegetation relationship. Geographic shifts in precipitation and air temperature patterns corresponded with vegetation shifts from grass dominance to shrub dominance. The soil physical and chemical properties (i.e., soil structure, bulk density, sodium absorption ratio, etc.), landscape, and microrelief have an effect on soil hydraulic conductivity and plant root size, abundance, and distribution within soil profile. The effects are evident in the variation in kind, amount, and proportion of vascular plants and associated cryptogam communities within similar climate regimes.

Soil taxonomic criteria provide a starting point for defining and extrapolating soil-climate-plant relationships. However, interactions with other environmental factors, more specific soil characteristics, and temporal influences also are needed.

Introduction

The areas in northern and eastern Montana are generally described as a grama-needlegrass-wheatgrass (*Boutelona-Stipa-Agropyron*) potential vegetation type in the central and eastern grasslands by (Küchler, 1964). The remaining areas in western Montana, Idaho, and Wyoming lie within the sagebrush-steppe (*Artemisia-Agropyron*) potential vegetation type of the western shrub and grasslands. The shift from a grassland potential vegetation to the sagebrush-steppe vegetation probably is tied more to timing and distribution of effective precipitation in relationship to growing season (Stoddart et al., 1975) than to soils.

The climographs (Kormondy, 1969) depicted in Figure 1a-e reflect the mean monthly precipitation and temperature patterns for Select National Oceanic and Atmospheric Administration (NOAA) Weather Stations that are representative of the tour sites. The precipitation that falls when the air temperature is below 32 degrees F. generally is assumed to be at a time when the soil is frozen and in the form of snow. Where the area is windswept and the snow is blown off the soil surface, soil moisture recharge from precipitation occurring during this period is considered to be low. Depending upon the site location, this can occur over a period of three to 6 months. In areas where snowpack accumulations are common, the reverse is also true. The importance of timing, amount, and distribution pattern of precipitation when the soils are not frozen is reflected in the efficiency of soil moisture recharge and kinds of vegetation communities.

Within the grassland areas involved, the dominant plant species are relatively shallow-rooted. Much of the moisture comes during the growing season but does not result in deep soil percolation. The different range sites occurring within these areas therefore can be determined satisfactorily, for the most part, by the soil characteristics in the upper layers that affect moisture relationships (texture, structure, salinity, etc.) and by effective precipitation.

Many of the grass species are constant between range sites. The relatively gentle topography contributes to rather broad climatic and

[1]G.J. Staidl and S.G. Leonard are soil scientist, USDA, SCS, and range scientist, USDI, BLM, respectively, National Soil-Range Team, Univ. of Nevada Reno, 1000 Valley Road, Reno, NV 89512.

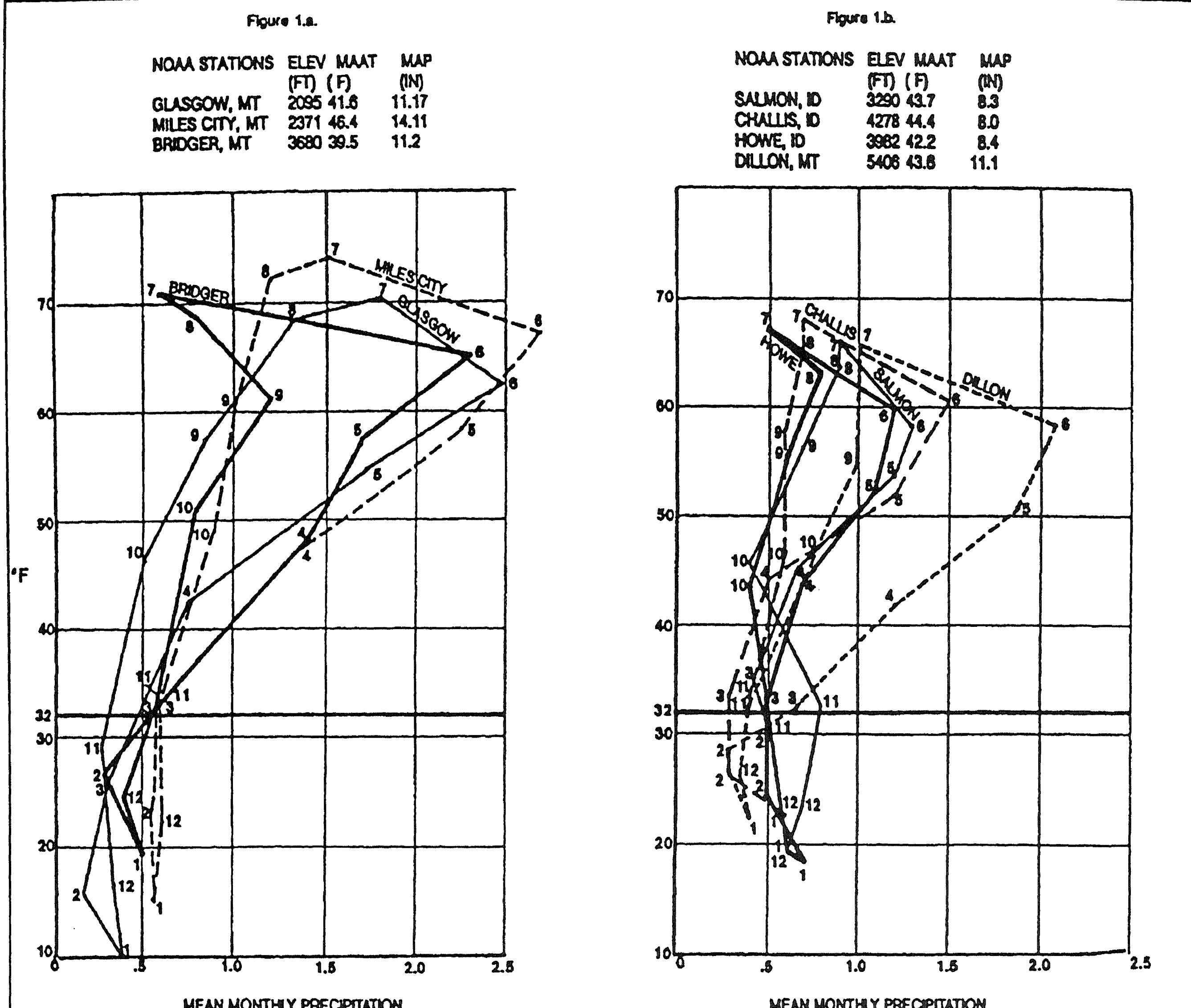
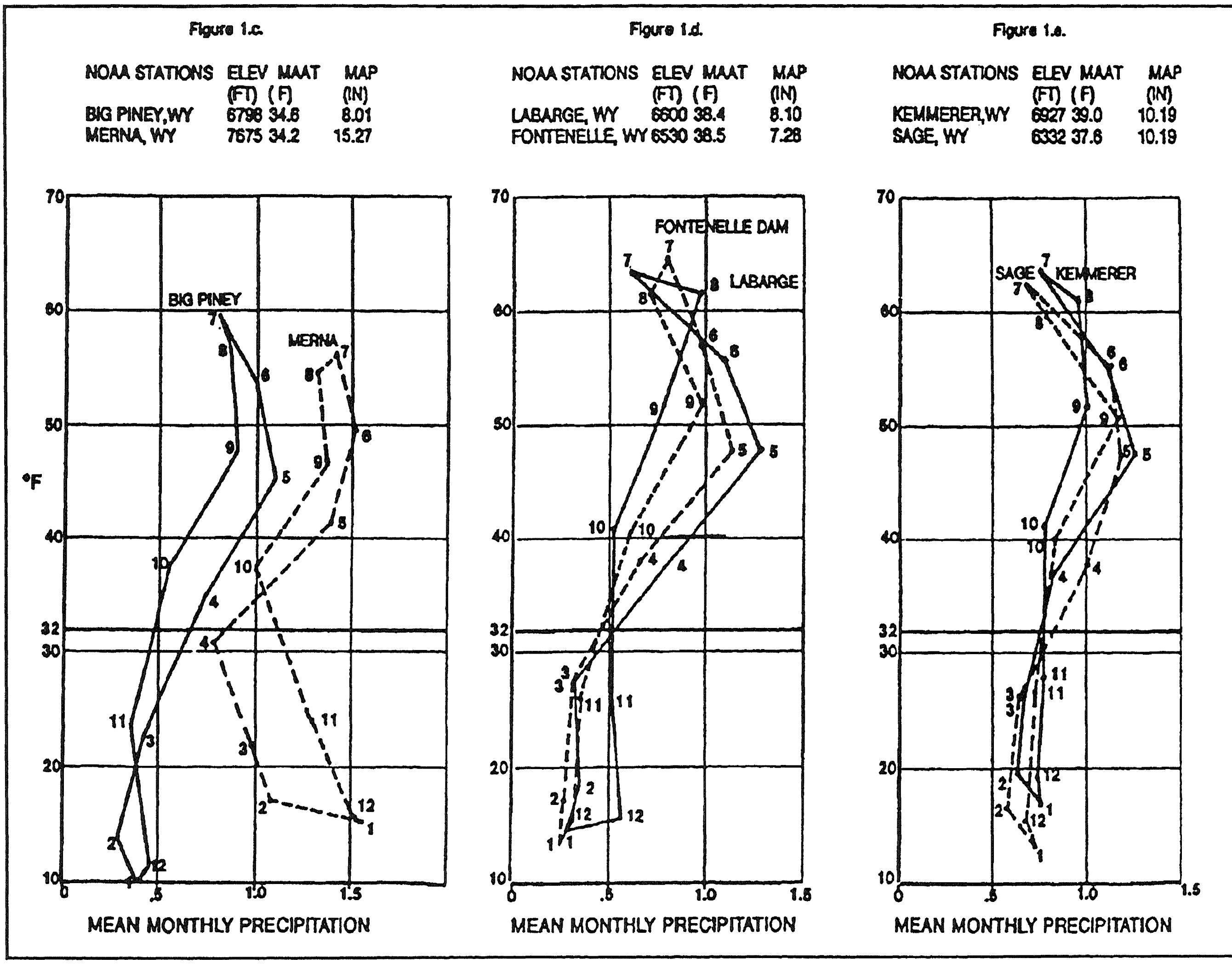


Figure 1. Climographs of representative NOAA Stations for each soil site stop along the tour. The vertical axis is temperature (°F); the horizontal axis is precipitation (inches); numbers refer to months (1=January, 2=February, etc.). Data are long-term monthly averages.

vegetation gradients except where affected by microtopographic influences and mans management practices. The sites often are differentiated more by total potential annual production by air-dry-weight and the relative proportion of species within the climax plant community than by distinct species changes.

Figure 1.a. represents the moisture and temperature patterns for Glasgow, Miles City, and Bridger in eastern Montana. Note that the monthly precipitation exceeds .4 cm per month throughout the main growing season. This pattern, along with the timing and kind of storms, is most favorable to grass vegetation (Figure 2). Bridger Station (Warren site) has a dry-down period in July and August and, although grass dominates, shows a small increase in shrubs and a corresponding decrease in percent cover of grass.

The sagebrush steppe is considerably more complex. Topography ranges from broad intermountain basins to rather narrow mountain valleys and mountain foothills. Elevation ranges within this type are great (699m - 2432m) and sometimes locally steep. Soil parent material also is quite varied, including but not limited to lake sediments, sedimentary rocks, basalt, and granitics. The concept for most of the sagebrush steppe is that precipitation tends to be more in the form of winter snow



and spring rains. A review of available climate data indicates this is not necessarily so at the eastern reaches of this vegetation type, but it is more arid than the grassland type.

Western Montana and eastern Idaho (Figure 1.b.) reflect the patterns similar to those shown in Figure 1.a. but also show a drying trend. The Dillon area still reflects a dominance of grass cover (Figure 2), but the relative cover of shrubs is increasing. The monthly precipitation at this site remains at or above .4 cm per month throughout the majority of the growing season.

In eastern Idaho the moisture pattern is somewhat dryer than at Dillon, MT, with the peak effective precipitation occurring during May and June and a dry-down occurring in July through October. The change in precipitation patterns has reduced the extended availability of surface soil moisture for grasses common to eastern Montana. This has allowed adaptable shrubs, with both lateral and taproot root systems that use both surface soil and deeper subsoil moisture, to compete effectively with the grasses for available water during the growing season. The Leador ID area in Figure 2 reflects this trend with a near equal mix of shrubs and grass.

Western Wyoming (Figures 1.c., 1.d., and 1.e.) precipitation patterns, although dryer, are similar, except the peak for moisture is generally in May, with a general drying starting in June. With surface soil moisture conditions less favorable, the shrub communities (Figure 2) are more dominant, due to their competitive ability to use both surface and deeper soil moisture.

Vegetation communities are often distinct within the sagebrush steppe, particularly between the various sagebrush taxa. These various taxa have been used successfully as indicator species (in conjunction with productivity and/or associated species) for major soil properties, as well as for range site differentiation. Figure 3 (Hironaka et al., 1983) provides a conceptual relationship of sagebrush species and

subspecies with soil characteristics as observed in Idaho.

West (1979) provides some general relationships between elevation, soil moisture, and various sagebrush species in the Great Basin and Colorado Plateau physiographic regions (Figure 4). Other authors, such as Zamora and Tueller (1973), Passey et al. (1982), and Sasich and Nielson (1984), to name a few, have also discussed soil, plant, and climate relationships of sagebrush species or communities in different locations.

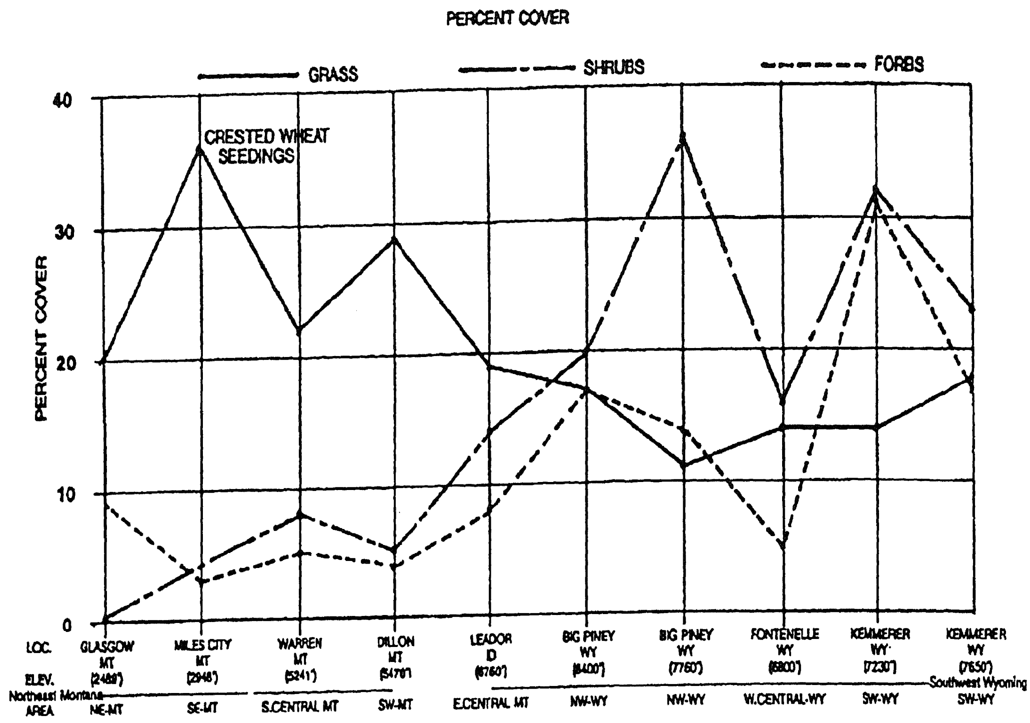


Figure 2. Total percent cover of grass, forb, and shrub vegetation for each tour site. Data was taken from Table 1.

While there are consistencies in some relationships, there are also apparent differences expressed by the above authors. The soil, plant, and climate relationships do vary slightly between geographic areas, and exceptions can be found within any geographic area. The geograpic differences can be accounted for, in part, by ecotypic variation within species or studies conducted at geographic extremes in the adaptive range of a species. More importantly, there are many compensating factors between soil, climate, topography, and other environmental factors that provide a physiologically similar environment for plant species (Leonard et al., 1988). *Soil Taxonomy* criteria may not always correspond with specific physiological responses of range plants.

Our preoccupation with soil, plant, and climate relationships and management implications associated with vascular plant communities has virtually ignored or at best barely recognized the "other" plant community component of most arid and semiarid ecosystems - the cryptogamic or nonvascular plant component.

Harper and Marble (1988) point out the importance of the cryptogamic plant community best, as follows:

Arid and semiarid rangelands commonly have blue-green algae, lichens, and mosses that often cover as much or more of the soil surface as vascular plants. Evidence suggests that most cryptogamic plants complement the effects of vascular plants relative to soil stability, water infiltration, and greater availability of nitrogen. Situations are documented in which cryptogamic covers (mainly blue-green algae) have improved establishment and growth of vascular plant seedlings.

The authors above also reference literature documenting certain potentially negative effects (i.e. allelopathy) associated with some cryptogams. It appears that there are nonvascular plant species that have both desirable and undesirable values or characteristics, much like, some species with the vascular plant community. It also appears that there are definite soil, climate, and topographic relationships affecting these species.

Methods and Materials

Each site identified as rangeland in the Idaho, Montana, and Wyoming part of the Aridisol tour were visited by the Soil Scientist and Range Scientist members of the National Soil-Range Team. Documentation available included the soil pedon descriptions and Soil Con-

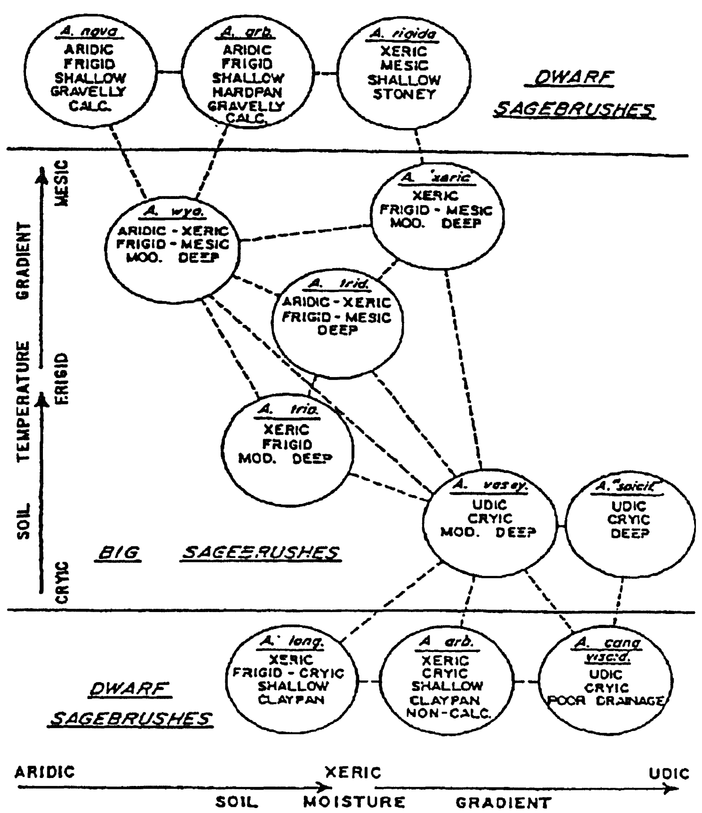


Figure 3. Conceptual relationship of sagebrush species and subspecies based on soil moisture, temperature and other soil characteristics.

servation Service National Soil Survey Laboratory characterization data. Supplemental support documentation, where available, included soil survey area manuscripts, field sheets, identification legends, geology reports, climate data, and vegetation data. The techniques and procedures used are the same as those outlined by the United States Department of Agriculture, Soil Conservation Service, in the *National Soils Handbook, Soil Survey Manual, Soil Taxonomy* and *National Range Handbook*.

Soil Profile Documentation

Each soil profile was examined in the field with an emphasis on the physical and chemical soil properties that would influence the present and potential plant community. The description of each pedon was compared with the soil profile. Particular attention was given to the size, abundance, distribution, and depth of plant roots. Any differences between the documentation and the observable soil properties that would influence the kind, amount, and proportion of plans were noted.

Vegetation Documentation

Production and composition air-dry-weight (ADW) was estimated using ocular estimates by SCS area range conservationists familiar with the local plant communities, except in Wyoming. In Wyoming, 1988 production and composition data are presented from double sampling measurements made by a BLM range conservationist. Cover estimates were obtained by the National Soil-Range Team in cooperation with SCS range conservationists, using the Daubenmire technique (USDI, 1985). Due to time constraints, only 20, 20 cm x 50 cm plots were estimated at each site, rather than the usual 40-50 plots. Only major plant species are listed, for the sake of brevity.

Cover estimates and composition summaries are attached to the soil data pertinent to each individual site and are located in the field guide used during the tour. Table 1 in this paper is a composite of the above grass, forb, and shrub vegetation data by percent for each site. Figure 2 graphically displays the data obtained from Table 1, to allow the user to compare, at each site, relevant soil-vegetation-climate parameters.

Except for the two sites near Kemmerer, WY, we attempted to quantify, as a cover component, the surface area affected by cryptogams. Cryptogamic surfaces were recognized by the roughened surface morphology or Type II surfaces (Eckert et al., 1986) associated with soil aggregates held together by mucilaginous secretions and/or the visual presence of lichen, moss, or clubmoss on the soil surface. During the vegetation data collection, no attempt was made to differentiate cryptogam species. Cover estimates are general because there was a great variation in thickness and degree of cryptogam development. The intent here is merely to emphasize the species presence and the fact that we are grossly neglecting a major plant component in our ecological analyses of soil, climate, and vegetation relationships.

At the Scobey site near Glasgow, MT, clubmoss (*Selaginella densa*) is a primary ground cover component. Clubmoss is considered very detrimental to vascular plant production (Dolan and Taylor, 1972) and establishment. The remainder of the sites all have varying degrees of algal and lichen development in and upon the soil surface (cryptogamic crusts). These crusts appear to have a favorable effect on seedling establishment or at least no detrimental effects.

Results and Discussion

Soil-Vegetation and Relationships

Montana - Scobey Soil

The vegetation components associated with the Scobey Series (NSSL ID#: 88PO849) produce an estimated 896 kg/ha (800 lbs./ac) with a cover dominated by grass, few shrubs, and intermediate amounts of forbs. Irrespective of the pedon description, the majority (common or many) of plant roots are concentrated within the upper 30 cm of soil and are generally exped below 13 cm, where the argillic horizon is assumed to start. Below 30 cm, the abundance of roots decreases dramatically when the whole horizon is taken into account, except along ped surfaces.

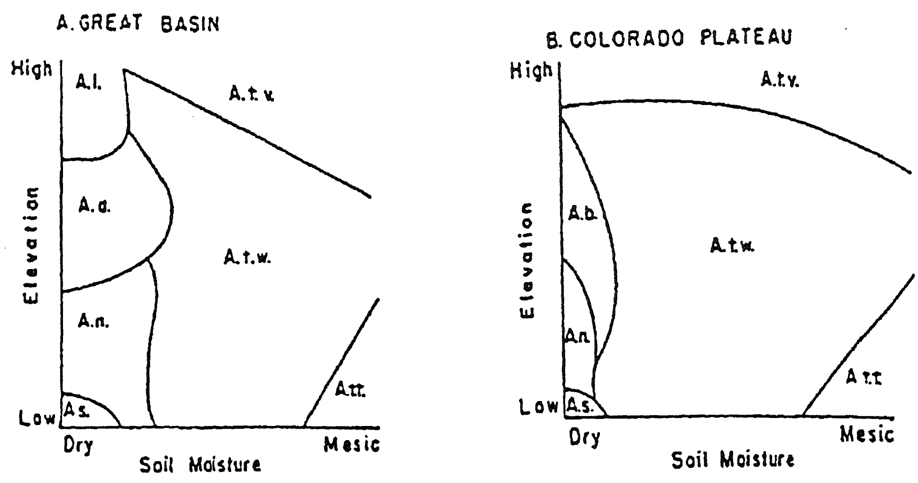


Figure 4. Ordinations' of major sagebrush taxa against gradients of elevation and effective moisture. The symbnols represent the first letters of the genus, species and subspecies names, therefore, A. = Artemisia; A.a.= A. Arbuscula; A.b.= A. bigelovii; A.l.= A. longiloba; A.n.= A. nova; A.s.= A. spinescens; A.t.t.= A. tridentata tridentata; A.t.v.= A. tridentata vaseyana; and A.t.w.= A. tridentata wyomingensis.

This soil lacks any apparent lithologic discontinuity in the upper soil profile that could restrict hydraulic conductivity. The climate pattern (Figure 1.a.) for Glasgow, MT, generally favors grass vegetation. A high amount of organic carbon is common to 30 cm. Evidence suggests that the depth of frequent wetting is approximately 30-61 cm, as the concentrations of finely divided lime, the sodium absorption ratio, and the pH greater than 8.5 collectively increase within this zone. Occasional deep wetting is evident from the increased electrical conductivity, reflecting soluble salt movement to depths greater than 79 cm. The weak, very fine platy structure at the soil surface at present is assumed to have a minimal effect on soil moisture infiltration. Bulk density and soil structure also plays an important role in affecting roots at depths below 13 cm.

Montana - Cambeth Soil

The vegetation cover components on the Cambeth Series (NSSL ID#: 88PO851) consist of a crested wheatgrass (*Agropyron cristatum*) seeding in a area that once was native range. The estimated production of crested wheatgrass is 952 kg/ha (850 lbs./ac). This site is located in an area on the upper third of the sideslope, with microrelief that is slightly convex. This type of landscape position is slightly more conducive to runoff. Infiltration of precipitation may be retarded further by the presence of a moderately strong or strongly developed, 0.5-1.0 cm thick vesicular crust.

Irrespective of the pedon description, plant roots range from many in the upper 14 cm to common to 24 cm and are reduced dramatically to few below 24 cm. This soil lacks any apparent lithologic discontinuity in the upper soil profile that could restrict hydraulic conductivity. The representative climate pattern (Figure 1.a.) for Miles City, MT, generally favors grass vegetation.

High amounts of organic carbon is common to 24 cm. Evidence suggests that the depth of frequent wetting is approximately 14-24 cm. The concentration of finely divided lime and the pH greater than 8.4 start to increase within this zone. Occasional deep wetting is evident from the high levels of finely divided lime, the sodium absorption ratio, and electrical conductivity reflecting soluble salt movement at depths greater than 24 cm.

Montana - Lonna Soil

The vegetation cover components on the Lonna Series (NSSL ID#: 88P0852) consist of a crested wheatgrass (*Agropyron cristatum*) seeding in a area that once was native range. The estimated production of crested wheatgrass is 15 kg/ha (1350 lbs./ac). This site is located in an area on the lower part of the sideslope com-

Table 1. Composite of Vegetation Cover, Composition and production Data Collected by Soil at Each Site.

General Site Locations	Total	Grasses		Forbs		Shrubs		Est. Prod.
	% Cover	% Cover	% Comp	% Cover	% Comp	% Cover	% Comp	kg/ha
Glasgow, MT (Marias) (721m) Site 1 Wheat Barley								
(Scobey)(757m) Site 2	29	20	86	9	8	T	2	896
Miles City, MT (Cambeth)(Convex)(896m) Site 1 Crested Wheat	35	32		3		—		952
(Lonna) (Concave)(896m) Site 2 Crested Wheat	48	46		2		—		1512
Warren, MT (Stormitt) (1593m) Site 1	35	22	83	5	4	8	13	392
Dillan, MT (Crago) (1663m) Site 1	38	29	87	4	10	5	5	336
Leador, ID (Arbus)(2116m) Site 1	40	18	45	8	5	14	50	336
(Not designated)(2116m) Site 2	47	20	40	5	8	22	53	672
Big Piney, WY (Chedsey tax)(2554m) Site 1	54	17	15	17	14	20	71	739
(Ashuelog var)(2359m) Site 2	61	11	16	14	11	36	73	649
Fontenelle Dam, WY (Tresano var)(2049m) Site 1	35	14	47	5	10	16	43	554
(Tresano var)(2073m) Site 2	29	5	19	10	23	14	58	448
Kemmerer, WY (Luhon var)(2326m) Site 1	58	18	37	17	13	23	50	784
(Iyers var)(2198m) Site 2	77	14	15	31	6	32	79	739

monly called the footslope and extends to the toeslope. The microrelief is slightly concave. This type of landscape position is slightly more conducive to receiving run-on water and additional sediment from higher surrounding areas. The possibility of a snowpack on this soil also could be a consideration. Reduction of infiltration characteristics may be only of a minor nature due to the weakly developed, 0.5-1.0 cm thick vesicular crust at the surface.

The plant roots identified in the soil profile range from many in the upper 31 cm to common to 100 cm, and few below 100 cm. This soil lacks any apparent lithologic discontinuity in the upper soil profile that could restrict hydrologic conductivity. The representative climate pattern (Figure 1.a.) for Miles City, MT, generally favors grass vegetation.

High amounts of organic carbon is common to 31 cm. Evidence suggests that the depth of frequent wetting is approximately 31-45 cm. The concentration of finely divided lime, the sodium absorption ratio, and electrical conductivity increase within this zone. Occasional deep wetting is evident from the high levels of the sodium absorption ratio, electrical conductivity, and pH at depths greater than 45 cm.

The concave position on the landscape allows for increased opportunities to receive run-on water and increased infiltration due to a lack of a strongly developed vesicular crust. This is reflected by the differences in several soil characteristics such as the depth of wetting and vegetation production between this soil and the associated Cambeth soil.

<u>Montana-Stormitt Soil</u>

The vegetation associated with the Stormitt Series (SSL ID#: 88P0853) produces an estimated 392 kg/ha (350 lbs./ac) with a cover dominated by grass, few forbs, and slightly higher amounts of shrubs. The microrelief for the area is smooth to slightly concave. Infiltration may be somewhat reduced by the presence of a less than 0.5 cm thick, weakly developed vesicular crust.

Review of the soil profile noted that the majority of plant roots are concentrated in the upper 33 cm and the roots are common to 48 cm and few below 48 cm. This soil has a lithologic discontinuity at approximately 48 cm that will affect hydraulic conductivity by restricting water movement.

The representative climate patterns (Figure 1.a.) for Bridger, MT, although seasonally warmer with a July-August dry-down, favor grass vegetation but at somewhat reduced production levels. Moderate amounts of organic carbon are common to 33 cm. Evidence suggests that the depth of frequent wetting is approximately 33-48 cm. The concentration of finely divided lime and the pH increase within this zone. Occasional deep wetting may be inferred

from the slight increase of electrical conductivity, reflecting soluble salt movement to depths greater than 80 cm. The location and concentration of plant roots reflect the above soil moisture relationships.

Montana - Crago Soil

The vegetation associated with the Crago Series (SSL ID#: 88P0854) produces an estimated 336 kg/ha (300 lbs./ac) with a cover dominated by grass and including a few forbs and shrubs. The low production may be due in part to plant species composition, resulting in the fair condition class.

Infiltration characteristics of this soil may be reduced by the presence of a 1-2 cm thick, moderately strong vesicular crust. Review of the soil profile noted that the majority of plant roots are concentrated in the upper 14 cm, and roots are common to 27 cm, common exped and along weak fracture plains to 83 cm, and none below 83 cm.

This soil has a lithologic discontinuity at approximately 27 cm that will have some affect on hydrologic conductivity by restricting water movement. The representative climate pattern (Figure 1.b.) for Dillon, MT, is seasonally cooler and somewhat dryer than Glasglow and Miles City. The pattern continues to favor grass vegetation but at reduced production levels.

High amounts of organic carbon are common to 27 cm. Evidence suggests that the depth of frequent wetting is approximately 27-55 cm, as the concentration of finely divided lime and the pH increased substantially within this zone. Occasional deep wetting is evident from the increased electrical conductivity, reflecting soluble salt movement and the increase in the sodium absorption ratio at depths greater than 55 cm. The location and concentration of plant roots reflect the above soil moisture relationships quite well.

Idaho-Arbus Soil

The vegetation associated with the Arbus Series (NSSL ID#: 88P0848) produces an estimated 336 kg/ha (300 lbs./ac), with a cover dominated by approximately equal amounts of grass and shrubs with few forbs. The weak or very weakly developed, 0.5-1.5 cm thick, vesicular crust at the soil surface is assumed to have a minimal effect on infiltration characteristics.

Review of the soil profile noted that the majority of plant roots are concentrated in the upper 23 cm, and roots are common on rock fragment surfaces and along weak fracture plains to 76 cm and generally no below 76 cm. This soil has a lithologic discontinuity at approximately 23 cm that will have some affect on hydrologic conductivity by restricting water movement.

The representative climate patterns (Figure 1.b.) for Challis, Howe, and Salmon, ID are seasonally cooler and dryer than Dillon, Glasgow, and Miles City, MT. The seasonally dryer pattern, with precipitation peaking in June then decreasing throughout the remaining months, tends to shift the vegetation communities towards shrub domination. These low sagebrush (*Artemisia arbuscula*) communities are highly competitive for any available soil moisture throughout most of the growing season. Low sagebrush is highly adapted to conditions ranging from saturation as a result of a perched water early in the growing season, to extreme dryness later in the summer, or season-long droughtiness associated with wind swept ridges, etc.

High amounts of organic carbon are common to 33 cm, with lesser amounts to 43 cm. Evidence of frequent wetting is approximately 23-33 cm. The concentration of finely divided lime and the pH increase within this zone. Occasional deep wetting is noted from increases in electrical conductivity and sodium absorption ratio at depths greater than 43 cm. The location and concentration of plant roots reflect the above relationships quite well.

Idaho-unnamed Soil

The vegetation associated with this unnamed soil (NSSL ID #: 88P0057) produces an estimated 672 kg/ha (600 lbs./ac), with a cover dominated by approximately equal amounts of grass, and shrubs, with few forbs. The soil shows ample evidence of numerous burrows from past animal activity, soil material scattered about the soil surface, and 5-10 cm in diameter krotovinas at depths ranging from about 30-70 cm. The high nitrate concentration below 42 cm also may be the end result of animal activity. Infiltration characteristics are assumed to be minimally affected because of a very weakly developed vesicular crust that ranges up to 1 cm.

Review of the soil profile noted that the majority of plant roots are concentrated in the upper 61 cm., are common to 100 cm, and few below 100 cm. This soil has a lithologic discontinuity at approximately 100 cm that may have some affect on hydrologic conductivity, assuming soil moisture penetrates to this depth.

The representative climate patterns (Figure 1.b.) for Challis, Howe, and Salmon, ID, are seasonally cooler and dryer than Dillon, Glasgow, and Miles City, MT. The seasonally dryer pattern, with precipitation peaking in June and then decreasing throughout the remaining months, tends to shift the vegetation communities towards shrub domination. Big sagebrush (*Artemisia tridentata*) communities are highly competitive in using both shallow and deep soil moisture throughout most of the growing season.

High amounts of organic carbon are common to 61 cm. Evidence of frequent or occasional wetting may by masked due to animal activity. However, wetting max be inferred from the electrical conductivity and sodium absorption levels increasing at 42 cm and reaching the highest levels at 61-100 cm. Finely divided lime is fairly evenly distributed to 100 cm and increases substantially along with pH below depths of 100 cm. This also could reflect previous early Holocene or late Wisconsin Pluvial climates.

<u>Wyoming - Chedsey Soil</u>

The vegetation associated with the Chedsey Taxadjunct Series (NSSL ID #: 88P0856) produces an estimated 739 kg/ha (660 lbs./ac), with a cover dominated by near equal amounts of grass, forbs, and shrubs, with shrubs comprised mainly of low sagebrush (*Artemisia arbuscula*) being slightly higher in proportion.

The thin vesicular crust on the soil surface is assumed to have some impact on infiltration characteristics. Review of the soil profile noted that the majority of plant roots are concentrated in the upper 23 cm, common to 45 cm, and few below. This soil lacks any apparent lithologic discontinuity in the upper soil profile that could restrict hydraulic conductivity.

The climatic pattern (Figure 1.c.) for Merna, Wyoming, generally represents the site. It is seasonally cooler and dryer than the Montana sites, but is slightly more moist than the Idaho sites during the growing season. The expected moisture distribution pattern is conducive to leaching and deeper soil water availability through the short growing season. The higher production and near equal amounts of grass, forbs, and shrubs reflect this climate pattern.

High amounts of organic carbon are common to 23 cm, with slightly lesser amounts to 45 cm. Sufficient seasonal moisture is available to leach carbonates and reduce the base saturation above 10 cm. Evidence suggests that the depth of frequent wetting is approximately 45-74 cm, as the concentration of pedogenic finely divided lime increases substantially. Occasional deep wetting is evident from the increased electrical conductivity and the pH at depths greater than 74 cm. Plant root penetration also ia reflected in the changes to bulk density at 23 to 45 cm and below 45 cm. The high bulk density also affects other soil properties that will affect plants, such as total pore space and available water capacity.

<u>Wyoming - Ashuelog Variant Soil</u>

The vegetation components associated with the Ashuelog Variant (NSSL ID#: 88P0855) produce an estimated 649 kg/ha (580 lbs./ac), with a cover dominated by shrubs and including smaller amounts of grass and forbs. The presence of a vesicular crust was not evaluated at this site. Review of the soil profile noted that the majority of plant roots were concentrated in the upper 33 cm, with few roots from 33 to 66 cm. Further root and soil moisture penetration is restricted at depths below 66 cm by a petrocalcic horizon.

The climatic pattern (Figure 1.c.) for Big Piney, Wyoming, generally represents the site. It is cooler and dryer than the Merna, Wyoming, Idaho, and Montana sites. The seasonally dryer pattern with precipitation peaking in April and May, with a dry-down during the remainder of the growing season, tends to shift the vegetation community toward shrub domination. The black sagebrush-early sagebrush (*Artemisia nova - Artemisia longiloba*) communities are highly competitive for any available soil moisture throughout most of the growing season.

High amounts of organic carbon are common to 33 cm with lesser amounts to 66 cm. The depth of frequent and occasional wetting is difficult to determine in this profile. The zone from 33-66 cm may be considered as a degrading petrocalcic horizon or as a newly forming calcic horizon, as this is the major zone of finely divided lime accumulation. Either case could restrict roots. The presence of few roots within this layer also may indicate a lack of frequent deeper moisture penetration restricting major plant use of the soil to the upper 33 cm.

<u>Wyoming - Tresano Soil</u>

The vegetation components associated with the Tresano Variant (NSSL ID#: 88P0858) produces an estimated 554 kg/ha (495 lbs./ac), with a cover dominated by shrubs, and including slightly lower amounts of grass and few forbs. Infiltration characteristics of this soil may be reduced by the presence of a 1-3 cm thick vesicu-

lar crust. Review of the soil profile noted that the majority of plant roots were concentrated in the upper 54 cm with few or no roots occurring below.

The climate pattern (Figure 1.d.) for Fontenelle Dam and LaBarge, Wyoming, generally represents the site. It is slightly warmer than the Big Piney and Merna, Wyomging, sites and is cooler and dryer than the Idaho and Montana sites. The seasonally dryer pattern, with precipitation peaking in April and May with a drydown during the remainder of the growing season, tends to shift the vegetation communities toward shrub dominance. The Wyoming big sagebrush (*Artemisia tridentata* ssp. *Wyomingensis*) communities are highly competitive in using both shallow and deep soil moisture throughout most of the growing season.

Moderate amounts of organic carbon are common to 25 cm, with lesser amounts to 54 cm. Evidence suggests that the depth of frequent wetting is approximately 25-54 cm, as the concentration of pedogenic finely divided lime and the pH increase substantially. Occasional deep wetting is noted from the increase in electrical conductivity reflecting soluble salt movement at depths greater than 79 cm. The sodium absorption ratio, which starts to increase at 25 cm, also reaches its highest levels below a depth of 79 cm.

Wyoming - Tresano Variant Soil

The vegetation components associated with the Tresano Variant (NSSL ID#: 88P0861) produces an estimated 448 kg/ha (400 lbs./ac), with a cover dominated by shrubs, and including few grasses and intermediate amounts of forbs. Infiltration characteristics of this soil may be reduced by the presence of a 1-3 cm thick vesicular crust. Review of the soil profile noted that the majority of plant roots are concentrated in the upper 26 cm, with the total amount dropping off with depth, as they are generally confined to exped surfaces.

The climate pattern (Figure 1.d.) for Fontenelle Dam and LaBarge, Wyoming, generally represents the site. It is slightly warmer than the Big Piney and Merna, Wyoming, sites and is cooler and dryer than the Idaho and Montana sites. The seasonally dryer pattern, with precipitation peaking in April and May with a drydown during the remainder of the growing season, tends to shift the vegetation communities toward shrub dominance. The Wyoming big sagebrush (*Artemisia tridentata* spp. *Wyomingensis*) communities are highly competitive in using both shallow and deep soil moisture throughout most of the growing season.

Moderate amounts of organic carbon are common to 18 cm, with lesser amounts to 26 cm. Evidence suggests that the depth of frequent wetting is approximately 18-26 cm, as the concentration of pedogenic finely divided lime and the pH increase substantially. Occasional deep wetting is evident from the increased electrical conductivity reflecting soluble salt movement at depths greater than 80 cm. The sodium absorption ratio which starts to increase at 26 cm, also reaches its highest levels below a depth of 80 cm. The plant rooting characteristics are reflected in the above relationships.

Wyoming - Iyers Variant Soil

The vegetation components associated with the Iyers Variant (NSSL ID#: 88P0860) produce an estimated 739 kg/ha (660 lbs./ac), with a cover dominated by a near equal amounts of shrubs and forbs and lower amounts of grass. This site is located in an area on the lower part of the sideslope commonly called the footslope, extending to the toeslope. The microrelief is slightly concave. This type of landscape position is more conducive to receiving run-on water in addition to the possibility of a snowpack accumulation. Infiltration characteristics of this soil may be reduced by the presence of a thin vesicular crust and clay textures. Reduced infiltration may be offset by the natural tendency of this high clay soil to form deep wide cracks when partially dry, allowing the soil profile to remoisten from below.

Review of the soil profile noted that the majority of plant roots are concentrated in the upper 25-30 cm. Smaller numbers of roots that generally occur along ped faces and cracks extend to deeper depths.

The climate patterns (Figure 1.e.) for Kemmerer and Sage, Wyoming, generally represent the site. They are seasonally the same as the other Wyoming sites and are cooler and dryer than the Idaho and Montana sites. The peak precipitation months are April and May, with a dry-down period during the remainder of the growing season. This pattern is more oriented to shrub dominance, as the early sagebrush (*Artemisia longiloba*) communities are highly competitive for available soil moisture throughout the growing season.

High amounts of organic carbon are common to 13 cm, with moderate amounts to 90 cm. Evidence suggests that the depth of wetting is approximately 45 to 90 cm, as the electrical con-

ductivity increases, reflecting soluble salt accumulation. The shrink-swell characteristic of this soil is an important feature to consider, as it affects kinds of plants and their natural rooting habitat.

Wyoming-Luhon Soil

The vegetation components associated with the Luhon Variant (NSSL ID#: 88P0859) produce an estimated 784 kg/ha (700 lbs./ac), with a cover dominated by shrubs and including slightly lower proportions of grass and forbs. Review of the soil profile noted that the majority of plant roots are concentrated in the upper 58 cm, with few roots along ped surfaces extending to deeper depths.

The climate data (Figure 1.e.) for Kemmerer and Sage, Wyoming, are the nearest data for the site. However, because of the elevational difference, the climatic pattern from the Merna site (Figure 1.c.) may be more representative. They are seasonally the same as the other Wyoming sites and are cooler and dryer than the Idaho and Montana sites. The peak precipitation months are April and May, with a dry-down period during the remainder of the growing season.

This pattern is more oriented to a shrub dominance, as the early sagebrush (*Artemisia longiloba*) communities are highly competitive for available soil moisture throughout the growing season.

High amounts of organic carbon are common to 18 cm, with moderately high amounts to 107 cm. Evidence suggests that the depth of frequent wetting is approximately 18 to 58 cm, as the pH and the concentration of finely divided lime increase substantially. Occasional deep wetting is evident from the decreased amount of roots occurring on ped surfaces at depths greater than 58 cm.

Summary and Conclusions

The natural plant community best adapted to a particular rangeland soil is dependent on the climate-soil interactions and the physiological adaptation of individual species to extracting moisture and nutrients at available times and locations within the soil profile. Climate-soil interactions of course, are, affected by geographic location and topographic position. The plants (nonvascular as well as vascular) also affect the climate-soil interactions through moisture interception, shading, evapotranspiration characteristics, and many other factors. There are also biological relationships (symbiosis, allelopathy, etc.) and temporal relationships (surface disturbance, climate fluctuations, use and management, etc.) that affect the community characteristics. Major relationships affecting potential plant communities are reflected by the physical and chemical properties of each soil. Variations in present productivity, composition, etc. often are reflected by temporal factors and biological interactions.

Individual and combined results of these complex interactions could be observed throughout the tour. General relations of ambient temperature and precipitation distribution with soil moisture availability were evident in the transition from grass to shrub dominance. Physiological plant adaptations were evident from the various sagebrush species and subspecies response observed in Idaho and Wyoming. Variations in productivity on any soil can be attributed in part to biological interactions (clubmoss effects on the Scobey soil) or temporal relationships (drought and range condition on the Crago soil). It is, therefore, unrealistic for *Soil Taxonomy* to address all combinations of these complex interactions.

The perception is that *Soil Taxonomy* should address natural plant communities to a sufficient extent to make general interpretations of soil-climate-plant relationships and guide the user toward more specific information needs. Soil moisture and temperature regimes are a concern in natural plant communities and particularly for Aridisols. These concerns are perhaps best expressed as questions.

Adjacent soils mapped in complex or association may have other classifications (i.e., mollisols in association with aridisols within the same map unit). Is soil moisture regime a function of climate, soil properties, or the combined effect on plant productivity?

If soil moisture regime (and temperature regime) relates to productivity, how much and under what conditions? Blaidsell (1958) reports a continuum of beginning growth temperatures for range plants starting near 0°C, well below the 5°C standard in *Soil Taxonomy*. This and other physiological adaptions (rooting characteristics) suggest a major portion of plant growth may be associated with shallower or sometimes deeper soil moisture availability rather than with what typically is considered as the soil moisture control section. Is additional soil moisture control section information and criteria for natural plant communities needed?

Many range plants, shrubs in particular, seem able to function physiologically at soil moisture tensions of 6.0 to 7.0 MPa (Caldwell, 1985). Is the soil really "dry" to these plants, or is it merely dry according to our agronomic perceptions of a 1.5 MPa tension wilting point?

Acknowledgements

Special thanks to the Idaho, Montana, and Wyoming Soil Scientists and Range Conservationists of the Bureau of Land Management and Soil Conservation Service who provided assistance and information for this report.

References

Blaisdell, J.P. 1958. Seasonal development of native plants on the upper Snake River plains and their relationship to certain climatic factors. USDA Tech. Bull. 1190:68.

Caldwell, M. 1985. Cold desert. *In*: Physiological ecology of North American plant communites. New York, London, Chapman and Hall.

Dolan, J.J. and J.E. Taylor. 1972. Residual effects of range on dense clubmoss and associated vegetation. Range Management. 25:32-37.

Harper, K.T. and J.R. Marble. 1988. A role for nonvascular plants in management of arid and semiarid rangelands. *In*: Tueller, P.T. (ed.) Vegetation science applications for rangeland analysis and management. Kluwer Academic Publishers. Dordrecht. Pp. 135-169.

Hironaka, M., M.A. Fosberg, and A.H. Winward. 1983. Sagebrush-grass habitat types of southern Idaho. Forest, Wildlife and Range Experiment Sta. Bull. 35. University of Idaho. Moscow, ID. 44 p.

Kormondy, E.J. 1969. Concepts of ecology. In: McElroy, W.D. and Swanson, C.P. (Ed.) Concepts of modern biology series. Prentice-Hall, Inc. Publishers. 209 p.

Küchler, A.W. 1964. Potential natural vegetation of the conterminous United States. Special Publication No. 36. American Geographical Society. New York.

Leonard, S.G., R.L. Miles and P.T. Tueller. 1988. Vegetation-soil relationships on arid and semiarid rangelands. *In*: Tueller, P.T. (ed.) Vegetation science applications for rangeland analysis and management. Kluwer Academic Publishers. Dordrecht. Pp. 225-252.

Passey, H.B., V.K. Hugie, E.W. Williams and D.E. Ball. 1982. Relationships between soil, plant community and climate on rangelands of the intermountain west. Tech. Bull. No. 1669. USDA, SCS. Washington, D.C.

Sasich, J. and G.A. Nielsen. 1984. Indicators of soil climate used by Montana soil classifiers: a survey. Montana Ag. Exp. St. Special Report 4. Montana State University, Bozeman. 45 p.

Stoddart, L.A., A.D. Smith and T.W. Box. 1975. Range Management. McGraw-Hill, Inc. New York. 532 p.

USDA. 1975. Soil Survey Manual. U.S. Dept. Agric. Handb. 18, 503 pp., Illus. (Supplements Replacing pp. 173-188, Issued May 1962)

USDA. 1975. Soil Taxonomy: a basic system of soil classification for making and interpretating soil surveys. Soil Conservation Service, U.S. Dept. of Agric. Handbook 436, 754 pp., Illus.

USDI. 1985. Rangeland monitoring trend studies. Technical Ref. 4400-4. USDI, Bureau of Land Management. Denver Service Center. Denver, CO. 130 p.

West, N.E. 1979. Basic synecological relationships of sagebrush-dominated lands in the Great Basin and the Colorado Plateau. In: Anon. The sagebrush ecosystem: A symposium. Utah State University College of Natural Resources. Logan, Utah.

Zamora, B. and P.T. Tueller. 1973. *Artemisia arbuscula, A. longiloba*, and *A. nova* habitat types in Northern Nevada. Great Basin Naturalist. 46:3 349-352.

Soil Management Research on the Clay Soils of Southwestern Ontario - A Review

J. A. Stone*

Abstract

Intensive row crop agriculture on the clay and clay loam soils of southwestern Ontario has resulted in surface soil structural deterioration and accompanying decreases in productivity and increases in nonpoint source surface water and groundwater contamination. Maintaining and/or improving soil quality depends upon understanding the extent and causes of structural deterioration and upon introducing a structure-improving component into row cropping systems. This paper presents a review of completed and ongoing research conducted on the clay soils of southwestern Ontario, addressing: 1) methods of measuring changes in structure; 2) the effect of long-term row crop production, tillage, and traffic on structural deterioration, and 3) methods for improving structure.

Introduction

Soil degradation, the depletion of productive capability, is a major concern across Canada (Anonymous, 1984). This paper focuses on the clay and clay loam soils in southwestern Ontario, generally defined as south and west of metropolitan Toronto (south of 43° 30' North latitude, Fig. 1.), which are major areas of corn (*Zea mays* L.) and soybean (*Glycine max*) production. On these soils, degradation is evidenced primarily by a gradual deterioration in the ability of the surface soil to retain an arrangement of solid and void space favorable for crop production. This is referred to as surface soil structural deterioration and is contributing to increased erosion, runoff, and compaction.

Surface soil structural deterioration is believed to be a result of moldboard plow based, intensive row crop production (Ketcheson, 1980; McKeague et al., 1987). The soil structure of the surface 30 cm appears to have deteriorated to an equilibrium level and is considered compact (McKeague et al., 1987). A contributing factor is the misuse of modern farm machinery which enables a farmer to till excessively and to till when the soil is too wet. In addition, the period when the soil is dry enough to till satisfactorily is reduced if the soil structure is poor, and a cycle is likely to develop in which the soil is increasingly damaged.

Soil structural deterioration not only lowers productivity but also contributes to contamination of surface water and groundwater, as well (Miller et al., 1988; Stone and Logan, 1989). Surface runoff of sediment-sorbed nutrients increases algal growth in lakes and streams. The application of increased amounts of agricultural chemicals, particularly nitrogen and pesticides, in response to losses in productivity, is resulting in the contamination of groundwater. The International Joint Commission addressed these concerns in the Revised Great Lakes Water Quality Agreement (Anonymous, 1987) and, as a result, major Federal and Provincial programs have been introduced which address soil degradation problems in southwestern Ontario.

Maintaining and/or improving soil and water quality depends upon understanding the extent and causes of structural deterioration and upon introducing a soil structure-improving component into row cropping systems. The objective of this paper is to present a review of completed and ongoing research conducted on the clay soils of southwestern Ontario, addressing 1) methods of measuring changes in soil structure, 2) the effect of long-term row crop production, tillage, and traffic on soil structural deterioration, and 3) methods for improving soil structure.

Soils and Climate

The majority of fine textured soils in southwestern Ontario were formed on till or lacustrine sediments. The major clay and clay loam soil series is Brookston (Evans and Cameron, 1983). It extends over 8000 km^2, primarily in extreme southwestern Ontario, and is one of the most productive soils in Ontario when tile drained.

The Brookston series soils generally are regarded as being poorly drained, fine textured, relatively stone free, and located on level to gently undulating topography. Family particle size is primarily fine clayey but ranges from 21 to

*Agriculture Canada Research Branch, Research Station, Harrow, Ontario N0R 1G0, Canada.

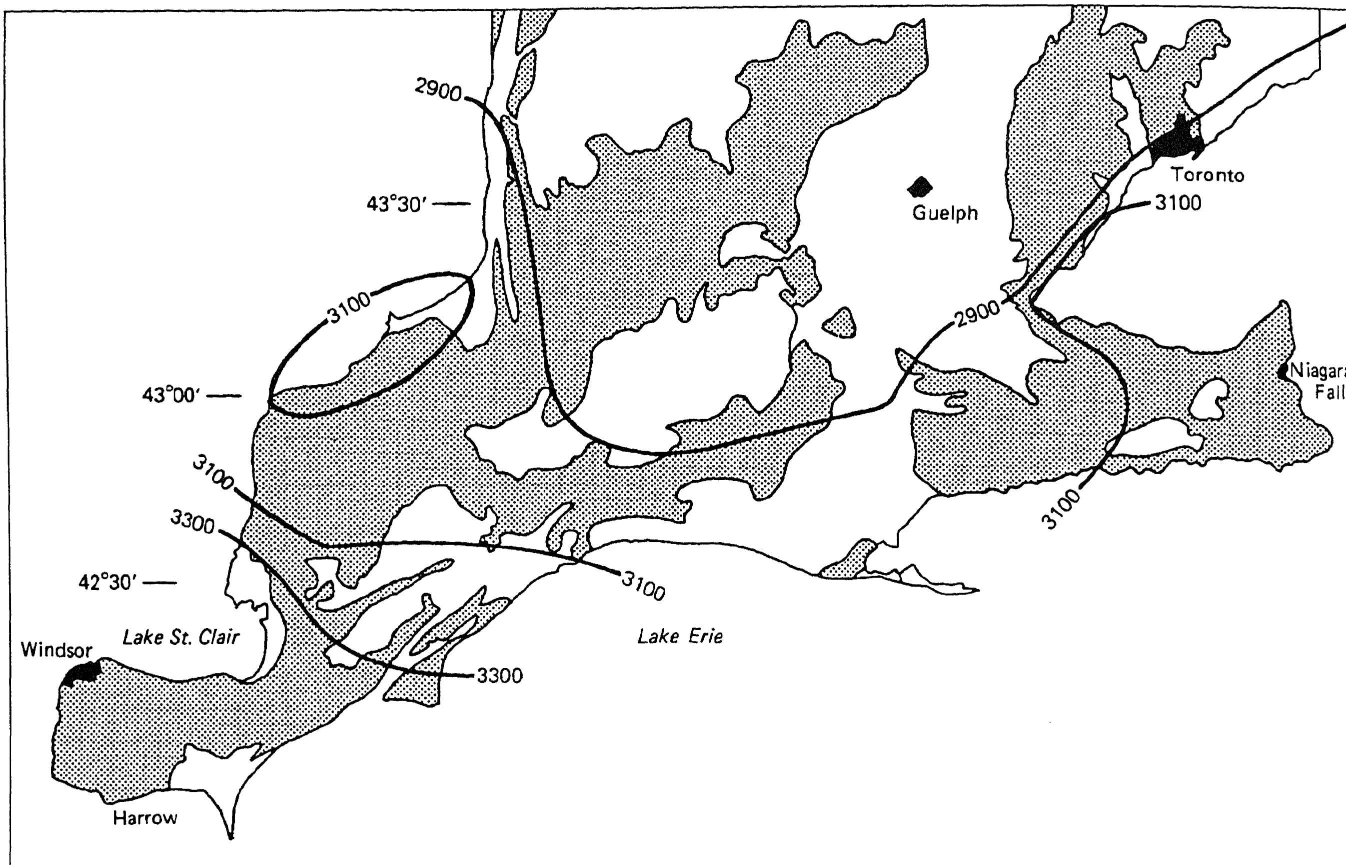


Fig. 1. Major areas of clay and clay loam soils in southwestern Ontario. The smooth curves delineate the heat units available for corn production.

65% (Anonymous, 1978). Texture is generally uniform to a depth of 1 to 1.5 m. Organic carbon in the surface horizon (Ap) averages 3.4%, except for extreme southwestern Ontario, where it averages 2.8%. Clay mineralogy consists predominately of illite, with subordinate amounts of chlorite, vermiculite, and hydroxy-interlayerd vermiculite (Evans and Cameron, 1983). Other major series of lacustrine origin, present in lesser amounts, are the very fine textured Haldimand and Lincoln soil series.

The climate of southwestern Ontario is influenced by the Great Lakes. The growing season ranges from approximately 135 to 175 frost free days (Wicklund and Richards, 1961). The heat units available for corn production range from approximately 2300 to 3600. Precipitation ranges from 800 to 1000 mm per year and is distributed fairly evenly throughout the year.

Measuring Changes in Clay Soil Structure

Quantifying soil structural deterioration requires a technique which is sensitive to changes but is relatively easy to apply. Changes in soil structure usually are assessed by performing a battery of tests, because no one technique is completely satisfactory.

Techniques that have worked best include soil core porosity and bulk density, and aggregate stability. Techniques that have not worked well, because of insufficient sensitivity and/or time required, include dry aggregate size distribution, aggregate shear strength, turbidity and instability index (Stone et al., 1985), and penetrometer resistance (Heslop et al., 1986).

This paper discusses core sampling, for bulk density and porosity determinations, and aggregate stability.

Bulk Density and Porosity

Core sampling with a typical hand held, hammer driven, double cylinder core sampler (Blake and Hartge, 1986) for the determination of dry bulk density, total porosity, and air-filled porosity has been used for many years. However, year-to-year fluctuations in bulk density and porosity, determined from soil cores, have been observed which are independent of soil and cropping treatments (Stone et al., 1985; Bolton et al., 1982).

Adjustment of long-term core data for volume changes resulting from variations in soil water content at the time of sampling were found to reduce, but did not eliminate, the year-to-year fluctuation (Stone and Wires, 1989). Stone (1989) compared hammer driven and hydraulic core sampling techniques and found that the core sampling technique significantly affected the overall mean values of bulk density, total porosity, and air-filled porosity. However, the sampling technique generally did not affect the ability of the core parameters to detect differences in soil structure at field soil water contents ranging from 12.6 - 23.8%.

These results (Stone, 1989; Stone and Wires, 1989) indicate that seasonal fluctuations in bulk density and porosity of Brookston clay loam occur independently of variations in water content at the time of sampling or in the sampling technique. Stone et al. (1985) and Bolton et al. (1981) found that seasonal fluctuations in bulk density and porosity related best to early spring rainfall. Bolton et al. (1981) concluded that heavy May-June rainfall contributed to a "slaking" effect which made the soil more susceptible to wheel traffic induced compaction. Horn (1988) explains that aggregate bulk density increases during cycles of wetting and drying, but varies with the extent of drying. Therefore, fluctuations in the water content of moist soil may promote rearrangement of particles and result in increases in bulk density.

Aggregate Stability

The most successful method for detecting changes in soil structure has been the determination of wet aggregate stability (WAS) on 0.5 to 2.0 mm air dry aggregates wetted by immersion (Kemper and Rosenau, 1986). Changes in WAS due to cropping and tillage have been shown (J. A. Stone, 1989, unpublished data; Stone and Heslop, 1987). However, in ongoing studies, WAS has been found to vary within the growing season, presumably due to soil moisture antecedent to sampling and the number of seasonal wetting and drying cycles (Stone, 1988). Research currently is underway to investigate the effect of water content on wet aggregate stability and dispersible clay (J. Caron, 1989, personal communication) and the relationship between aggregate stability and microbial biomass (J. A. Stone, 1989, personal communication).

Soil Structure and Row Crop Production

Long-term Corn Production

Long-term plots established on Brookston clay loam have provided information about the extent and effects of soil structural deterioration resulting from row crop production systems. McKeague et al. (1987) reported changes in soil structure resulting from long-term continuous and rotation corn in comparison with a never cultivated soil. The long-term corn plots (continuous and rotation) compared to the never cultivated plots had a massive as opposed to strong ped structure, lower macroporosity, isolated as opposed to interconnected pores, and greater bulk densities. They reported that the surface horizons of the rotation corn plots differed from those of the continuous corn plots in having more abundant biopores. However, below a depth of 30 cm, comparable horizons from all plots were similar in ped structure, bulk density, and water characteristic curves. There was no distinct plow pan or restriction in root distribution (Stone et al., 1987) resulting from long-term corn production.

Stone et al. (1985) summarized data for the period 1957-1982 on bulk density and porosity of Brookston clay loam soil from a long-term corn fertility experiment. They showed that the bulk density tended to slowly increase and porosity to slowly decrease with time in the 5 to 15 cm layer for both rotation and continuous corn. The rotation plots had lower bulk densities and higher porosities than the continuous corn plots. However, the fertility level had no effect on bulk density or porosity of the continuous corn plots.

The overall similar rates of deterioration of the continuous and rotation corn plots led Stone et al. (1985) to conclude that continuous corn by itself appeared to play a minor role in the deterioration of soil structure. By 1957, 4 years after the plots were established, dry bulk densities of all the plots were considerably higher than those of the never cultivated plot reported by McKeague et al. (1987). Compaction of the surface horizon apparently occurred prior to the monitoring of bulk density which began in 1957.

Bryant et al. (1987) analyzed tile flow data from long-term plots established on Brookston clay loam to determine the effect of cropping system on tile drain discharge. They found that

continuous corn and bluegrass generally contributed to a greater tile discharge volume than rotation corn, because of a longer flow period. For the bluegrass plots, this may be attributed to a more favorable soil structure and, therefore, more water storage in the surface profile. For the continuous corn plots, this may be a result of the compact soil structure and, therefore, ponding and longer wet periods.

Tillage and Traffic

Developing improved management practices to stabilize or improve soil structure requires information about the contribution of tillage and traffic to soil structural deterioration. Studies conducted on Brookston clay loam have provided conflicting information about the effect of tillage on soil structural deterioration.

Tillage Method

Stone and Heslop (1987) measured organic carbon, wet aggregate stability, and bulk density in a study comparing ridge, fall moldboard plow, and blade cultivator tillage in a corn - soybean - corn rotation. After 3 years, mid - season organic carbon ranked by tillage treatment was blade cultivator > ridge > moldboard plow, but the blade was not significantly different from ridge tillage. The values of wet aggregate stability were ranked as ridge > blade cultivator > moldboard plow.

There were no significant differences in bulk density between treatments. Increases in organic carbon and wet aggregate stability due to treatments were evident below the depth of tillage (20 to 30 cm). This suggests that improvements in soil structure resulting from these reduced tillage systems may be a result of root exudates and the byproducts of residue decomposition.

In another 3-year study conducted on Brookston clay loam (Stone et al., 1989), mid-season measurements of organic carbon, aggregate shear strength, and wet aggregate stability did not provide any consistent or meaningful differences between ridge, fall moldboard plow, or zero tillage (J. A. Stone, 1989, unpublished data). Corn - corn - soybean and soybean - soybean - corn rotations were compared at adjacent sites in this study.

Average organic carbon levels were lower at the site of this study than the site utilized by Stone and Heslop (1987). However, soil texture and wet aggregate stability were similar at both sites. Tile drain spacing was 12.2 m and 7.6 m for the sites utilized by Stone et al. (1989) and Stone and Heslop (1987), respectively. The slopes of the two sites were essentially equal (<0.5%), but surface drainage was somewhat restricted in the study of Stone et al.

The contrary results of these two studies suggest that improvements in soil structure resulting from reduced tillage on clay and clay loam soils may be enhanced by better drainage. They also suggest that, in the short-term, expected reductions in erosion and runoff resulting from reduced tillage are more likely induced by the physical effects of surface residue rather than from improvements in soil structure.

Wheel Traffic

Stone (1987) reported the results of a 3-year experiment conducted to determine the cumulative contribution of surface soil compaction to structural deterioration.

Fall vehicle compaction prior to moldboard plow tillage did not significantly contribute to surface soil structural deterioration. Fall compaction was presumably alleviated by fall plowing, freezing and thawing, spring secondary tillage, and wetting and drying cycles. Spring compaction generally contributed to structural deterioration; however, there were no cumulative detrimental effects of fall or spring compaction on soil structure.

It was concluded that the contribution of vehicle compaction, at soil moisture contents suitable for tillage, to the compact surface structure of the intensively cropped clay soils in southwestern Ontario, does not appear to be long-term and can be minimized by controlling spring wheel traffic. Similarly, Bolton and Aylesworth (1959) reported that the effects of excess tillage on pore space were small and that there was little or no association between porosity values and crop yield on this soil.

Although spring wheel traffic results in seasonal compaction and tillage is a major factor in the rate of drying in the spring, wheel traffic and tillage do not appear to be contributing to additional structural deterioration on clay soils in southwestern Ontario. This information is the basis for additional studies on the role of rhizosphere effects of grass and legume forages in improving soil structure.

Improving Soil Structure

The long-term productivity of the clay soils in southwestern Ontario appears to depend upon introducing a structure-improving component into the row cropping system. Grass and legume forages generally are accepted as the most effec-

tive means of improving soil structure. Recent studies have shown that forages may improve soil structure within one cropping season (Stone and Buttery, 1989; Stone, 1988; Angers and Mehuys, 1988).

Forages

Stone and Buttery (1989) reported that reed canary grass improved soil structure compared with other grass and legume forages, due to increased root mass. The frequency of VA mycorrhiza hyphae was not associated with improvements in structure, although it varied between forages. Drury et al. (1989) measured microbial biomass C and N under corn, soybeans, red clover, alfalfa, reed canary grass, orchard grass, and bare soil at monthly intervals during the third growing season on Brookston clay loam soil. They found that reed canary grass resulted in the greatest biomass C content and corn in the least. Varying the extent of the wetting and drying cycles in the bare soil did not affect biomass C or biomass N levels. Reed canary grass also showed the greatest improvement in simultaneous measurements of aggregate stability.

Microbial Factor

Fysun and Oaks (1989) have established that there is a microbial factor associated with legumes which dramatically increases the growth of corn under greenhouse conditions. Ongoing research (A. Oaks, 1989, personal communication) is being directed toward isolation and identification of the beneficial microorganisms.

Preliminary field studies carried out on Brookston clay loam indicate that inoculation of soils that have been under corn for 30 years leads to significant growth responses. Fysun and Oaks (1987) reported that planting corn in soil that has been under alfalfa leads to larger soil sheaths (soil adhering to roots) on seedling roots than when the corn seedlings have been grown in soil that has been under corn. The size of the sheath may relate to the detrimental effect of continuous corn on soil structure, and the simple measurement of this parameter may provide a useful assay of the degree of soil structural deterioration.

Intercropping

Probably the most economical method of incorporating the beneficial effects of forages into an intensive row crop production system is by interseeding the row crop with a forage, thereby providing surface cover, improvements in soil structure, and biologically fixed nitrogen. Considerable intercropping research was conducted in the 1950s, with limited success, due primarily to the lack of satisfactory weed control. Recently, researchers have successfully intercropped corn with a forage using chemical weed control (Scott et al., 1987; R. W. Sheard, 1988, personal communication).

Ongoing experiments on Brookston clay loam are designed to evaluate corn intercropped with legume and grass forages under conventional and conservation tillage, relative to improvements in soil structure while maintaining acceptable yields (J. A. Stone, 1989, personal communication). Related work is being conducted for a range of soil textures at the University of Guelph (B. D. Kay, 1989, personal communication).

Summary

Intensive row cropping of the clay and clay loam soils of southwestern Ontario has resulted in a gradual deterioration of the surface soil structure which is contributing to increased runoff, erosion, and compaction. Research is addressing: 1) methods of measuring changes in soil structure; 2) the effect of long-term row crop production, tillage, and traffic on soil structural deterioration, and 3) methods for improving soil structure.

Changes in structure have been assessed satisfactorily by determining bulk density, porosity, and aggregate stability. However, seasonal fluctuations exist which have not been fully explained. Ongoing research is investigating the role of seasonal wetting and drying cycles.

Evaluation of long-term corn plots has shown that the soil structure has deteriorated to an equilibrium level under continuous row cropping. Structural deterioration is restricted to the surface 30 cm and has not resulted in a distinct plow pan or restriction in root distribution. However, it results in a greater volume of discharge through tiles because of a longer flow period. Wheel traffic and tillage do not appear to be contributing to additional structural deterioration.

Long-term soil productivity appears to depend upon introducing the structure-improving attributes of forages into row cropping systems. Research has shown that forages differ in the ability to improve soil structure and indicates that there is a relationship between the microbial population associated with different forage species and the ability to improve soil structure. Ongoing research is being directed toward isola-

tion and identification of beneficial microorganisms. Corn - forage intercropping studies are underway to determine the feasibility of incorporating the beneficial effects of forages into an intensive row crop production system.

Literature Cited

Angers, D. A. and G. R. Mehuys. 1988. Short-term effects of cropping on macro-aggregation of a marine clay soil. Can. J. Soil Sci. 68:723-732.

Anonymous. 1978. The Canadian System of Soil Classification. Canada Soil Survey Committee. Pub. 1646, pg.115-118. Agriculture Canada Research Branch, Central Exp. Farm, Ottawa, Ont. K1A 0C6.

Anonymous. 1984. Soil at risk: Canada's eroding future. Published by the Standing Committee on Agriculture, Fisheries, and Forestry, Committees and Private Legislation Branch, The Senate of Canada, Ottawa, Ontario K1A 0A4.

Anonymous. 1987. Revised Great Lakes Water Quality Agreement of 1978. International Joint Commission, Great Lakes Regional Office, 100 Ouelette Ave., Windsor, Ont. N9A 6T3.

Blake, G. R. and K. H. Hartge. 1986. Bulk Density. *In* A. Klute (ed.) Methods of soil analysis. Part I. 2nd ed. American Society of Agronomy, Madison, WI.

Bolton, E. F. and J. W. Aylesworth. 1959. Effect of tillage traffic on certain physical properties and crop yield on Brookston clay soil. Can. J. Soil Sci. 39:98-102.

Bolton, E. F., V. A. Dirks, and M. M. McDonnell. 1981. Effect of fall and spring plowing at three depths on soil bulk density, porosity and moisture in Brookston clay. Can. Agric. Eng. 23:71-76.

Bolton, E. F., V. A. Dirks and M. M. McDonnell. 1982. The effect of drainage, rotation and fertilizer on corn yield, plant height, leaf nutrient composition and physical properties of Brookston clay soil in southwestern Ontario. Can. J. Soil Sci. 62:297-309.

Bryant, G. J., R. W. Irwin, and J. A. Stone. 1987. Tile drain discharge under different crops. Can. J. Agric. Eng. 29:117-122.

Drury, C. F., J. A. Stone, and W.I. Findlay. 1989. Seasonal changes in microbial biomass under grass and legume forages on a clay loam soil. (In press, copies available from C. F. Drury, Agriculture Canada Research Station, Harrow, Ont. N0R 1G0).

Evans, L. J. and Cameron, B. H. 1983. The Brookston series in southwestern Ontario: characteristics, classification and problems in defining a soil series. Can. J. Soil Sci. 63:339-352.

Fysun, A. and A. Oaks. 1989. Growth promotion of maize by legume soils. Plant and Soil (In press, copies available from A. Fysun, Dept. of Biology, McMaster Univ., Hamilton, Ont. L8S 4K1).

Fysun, A. and A. Oaks. 1987. Physical factors involved in the formation of soil sheaths on corn seedling roots. Can. J. Soil Sci. 67:591-600.

Heslop, L. C., J. A. Stone, and B. Compton. 1986. Portable loading frame for field penetrometer measurements. ASAE paper no. 86-1039. ASAE, 2950 Niles Rd., St. Joseph, MI 49085.

Horn, R. 1988. Strength of structured soils due to loading - a review of macro- and microscale processes; European aspects.*In* W. E. Larson, G. R. Blake, R. R. Allmaras, and S. C. Gupta (ed.) Mechanics and Related Processes in Structured Agricultural Soils. Kluwer Academic Publishers, Dordrecht. ISBN 0-7923-0342-3.

Irwin, R. W., G. J. Bryant, M. R. Toombs, and J. A. Stone. 1987. Evaluation of a drainage coefficient for a poorly drained soil. Trans. ASAE 30:1343-1346.

Kemper, W. D. and R. C. Rosenau. 1986. Aggregate stability and size distribution. *In* A. Klute (ed.) Methods of soil analysis. 2nd ed. Part 1. American Society of Agronomy, Madison, WI.

Ketcheson, J.W. 1980. Long-range effects of intensive cultivation and monoculture on the quality of southern Ontario soils. Can. J. Soil Sci. 60:403-410.

McKeague, J. A., C. A. Fox, J. A. Stone, and R. Protz. 1987. Effects of cropping system on structure of Brookston clay loam in long-term experimental plots at Woodslee, Ontario. Can. J. Soil Sci. 67:571-584.

Miller, M. H., Groenevelt, P. H. and Stonehouse, D. P. 1988. Stewardship of soil and water in the food production system. Notes on Agriculture. 22:5-12. Published by the Ontario Agricultural College, Univ. of Guelph, Guelph, Ont. N1G 2W1.

Scott, T. W., Mt. Pleasant, J., Burt, R. F., and Otis, D. J. 1987. Contributions of ground cover, dry matter, and nitrogen from intercrops and cover crops in a corn polyculture system. Agron. J. 79:792-798

Stone, J. A. 1987. Compaction and the surface structure of a poorly drained soil. Trans. ASAE 30:1370-1373.

Stone, J. A. 1988. Field crop and forage effects on seasonal values of structural properties of a poorly drained clay loam soil. *In* W. E. Larson, G. R. Blake, R. R. Allmaras, and S. C. Gupta (ed.) Mechanics and Related Processes in Structured Agricultural Soils. Kluwer Academic Publishers, Dordrecht. ISBN 0-7923-0342-3.

Stone, J. A. 1989. Comparison of hammer and hydraulic core samplers for soil bulk density and porosity determinations. ASAE/CSAE Paper no. 89-2191. ASAE, St. Joseph, MI.

Stone, J. A., N. H. E. Allen, and C. D. Grant. 1985. Corn fertility treatments and the surface structure of a poorly drained soil. Soil Sci. Soc. Am. J. 49:1001-1004.

Stone, J. A. and B. R. Buttery. 1989. Nine forages and the aggregation of a clay loam soil. Can J. Soil Sci. 69:165-169.

Stone, J. A. and L. C. Heslop. 1987. Blade cultivator, ridge, and moldboard plow tillage comparison on a poorly drained soil. Trans. ASAE 30:61-64.

Stone, J. A. and L. L. Logan (ed.). 1989. Agricultural Chemicals and Water Quality in Ontario. Proc. Workshop, Kitchener, Ont., Nov. 17-18, 1988. ISBN 0-921447-05-1. (Copies available from J. A. Stone, Agriculture Canada Res. Branch, Res. Sta., Harrow, Ont. N0R 1G0).

Stone, J. A., J. A. McKeague, and R. Protz. 1986. Corn root distribution in relation to long term rotations on a poorly drained clay loam soil. Can. J. Plant Sci. 67:231-234.

Stone, J. A., T. J. Vyn, H. D. Martin, and P. H. Groenevelt. 1989. Ridge-tillage and early-season soil moisture and temperature on a poorly drained soil. Can. J. Soil Sci. 69:181-186.

Stone, J. A. and K. C. Wires. 1989. Moisture content and soil core volume on Brookston clay loam. Can. J. Soil Sci. (In press, copies available from J. A. Stone, Agric. Canada Res. Branch, Res, Sta., Harrow, Ont. N0R 1G0).).

Wicklund, R.E., and Richards, N. R. 1961. The soil survey of Oxford County. Report No. 28 of the Ontario Soil Survey. Ontario Institute of Pedology, Guelph Agric. Centre., Box 1030, Guelph, Ont. N1H 6N1.

Answers to Questions

1. Larry Wilding: What are the problems in using the Guelph permeameter to characterize soil degradation?

First, I'd like to say that although the Guelph permeameter has not worked well as a measure of soil structural degradation for me on clay loam soil, I have not worked extensively with the technique. The major problems have been 1) boring a hole that does not intersect with a crack in the soil and 2) the amount of time required to obtain a measurement (because of the low hydraulic conductivity of these soils). I think that 1) can be overcome by conducting measurements when cracks are not present (in the spring or fall) and 2) by using a bank of permeameters. However, seasonal labour may not be available in the spring and the fall and using a bank of permeameters is expensive.

2. Larry Wilding: Have you explored with the application of micromorphology as a analytical tool to document structural degradation and changes in shape, size, and distribution of macroporosity?

Yes, macro- and microstructure were described and water desorption characteristics were measured for Brookston clay loam soil under different cropping systems. This information is presented in the following publication:

McKeague, J. A., C. A. Fox, J. A. Stone, and R. Protz. 1987. Effects of cropping system on structure of Brookston clay loam in long-term experimental plots at Woodslee, Ontario. Can. J. Soil Sci. 67:571-584.

Soil Classification Related Properties of Salt-Affected Soils

I. Szabolcs[1]

Abstract

Salt-affected soils which occur under different environmental conditions and have diverse morphological, chemical, physical, physico-chemical, and biological properties should be included in one big group, according to their common feature, the dominating influence of electrolytes on their formation and properties.

In the various soil classification systems, the names and places of salt-affected soils on the taxonomic levels are diverse, even though their morphological diagnostic features and their chemical, physical, and physico-chemical properties are very similar.

The following properties are significant for salt-affected soils in most classification systems:

1/ Morphology of the soil profile (presence or absence of diagnostic horizons).

2/ Significant physical properties (mainly for the Solonetz group).

3/ Chemical and physico-chemical properties:

/a/ Salt content, salt composition, and salt distribution in the profile and, in some cases, also in the ground water.

/b/ Exchangeable sodium percentage and sodium adsorption ratio.

/c/ pH conditions and the existence of sodium carbonate.

Most of the listed properties are included in the main international or national classification systems but on different levels with different limit values and sometimes with different interpretations.

Definition and Characteristics of Salt-Affected Soils

Salt-affected soils occur in more than a hundred counties, from above the polar circle to the equator. They occur under different environmental conditions and have diverse morphological, chemical, physical, physico-chemical, and biological properties.

Although the definition of salt-affected soils, and the taxonomic level at which they are classified, differ in the different countries and classification systems, all types, kinds, or varieties of these soils mainly are characterized by the presence and/or influence of water soluble salts in certain horizons or layers of the soil profile.

While the decisive influence of electrolytes is common in all salt-affected soils, their chemical composition and concentration vary, depending on local conditions. Consequently, different types of salt-affected soils develop with diverse properties.

Table 1, based on the above considerations, demonstrates a simple but practical grouping system for salt-affected soils. It classifies them in compliance with the types of electrolytes causing salinity and/or alkalinity.

The taxonomic units of Table 2 have been selected according to this paper's definition of salt-affected soils and their grouping in Table 1.

In the so-called genetic soil classification system which is common in the USSR (Kovda-Rozanov, 1988) and many other countries, Solonetz and Solonchak and sometimes Solod soils are high taxonomic units subdivided into types, subtypes, and varieties.

Table 1. Grouping of salt-affected soils

Electrolytes(s) causing salinity and/or alkalinity	Type of salt-affected soil	Environment	Main adverse effect on production	Method for reclamation
Sodium chloride and sulphate (in extreme cases - nitrate)	Saline soils	Arid and semi-arid	High osmotic pressure	Removal of excess salt (leaching)
Sodium ions capable of alkaline hydrolysis	Alkali soils	Semi-arid Semi-humid Humid	Alkali pH Effect on water physical soil properties	Lowering or neutralizing the high pH by chemical amendments
Magnesium ions	Magnesium soils	Semi-arid Semi-humid	Toxic effect, high osmotic pressure	Chemical amendments Leaching
Calcium ions (mainly $CaSO_4$)	Gypsiferous	Semi-arid Arid	Acidic pH, toxic effect	Alkaline amendments
Ferric and aluminum ions (mainly sulphates)	Acid sulphate soils	Sea shores, lagoons with heavy, sulphate containing sediments	Strongly acidic pH, toxic effect	Liming

[1]Research Institute for Soil Science and Agricultural Chemistry of the Hungarian Academy of Sciences, P.O.B. 35. H-1525 Budapest, Hungary.

Most of the other soil classification systems place the different salt-affected soils similarly to one or another of the three classification systems described above, (1. SZABOLCS, 1989; 2. Key to Soil Taxonomy, 1985; 3. Kovda and Rozanov, 1988), or in a manner which combines these systems' approaches.

The soil classification system in Canada is based on soil genetical principles but uses the terms and taxonomic structures of both the classification in the FAO/UNESCO World Soil Map and the US *Soil Taxonomy* (The Canadian System of Soil Classification, 1978).

The recent Australian soil classification principally is based, even more than the Canadian system, on traditional genetic terms and elements, but in this system a nearly complete correlation with the US *Soil Taxonomy* and with the Legend of the FAO/UNESCO World Map has been elaborated (Soils, an Australian viewpoint, 1983).

Discrepancies among classification systems' heirarchical placement of various salt-affected soils often make it difficult to find an approximate equivalent in one system of a type or variety defined in another system. However, the definitive role of electrolytes in the characterization of all salt-affected soils can facilitate the conversion of terms for such soils among the different classification systems.

Main Properties of Salt-Affected Soils Underlying Classification

Morphological and Physical Soil Properties Related to the Classification of Salt-affected Soils

In the three above-mentioned soil classification systems, and in almost all common soil classification systems, the main diagnostic morphological and physical feature for the characterization of Solonetz soils, Solodic soils, and the majority of alkali soils is the existence of a natric horizon that is used in the US *Soil Taxonomy* and the FAO/UNESCO Soil Map Legend. In the genetic and related soil classification systems, the criterion is the presence of nearly identical horizon, namely the horizon B.

The natric horizon is an argillic horizon which has prisms or, more commonly, columns in the subsurface, as well as blocky structures and tongues of an eluvial horizon. Other characteristics of natric horizons are mentioned in the next section of this paper.

Table 2. Salt-affected soil in the hierarchy of the US Soil Taxonomy

Order	Suborder	Great Group
Alfisols	Aqualfs	Natraqualfs
	Boralfs	Natriborals
	Udalfs	Natrudalfs
	Ustalfs	Natrustalfs
	Xeralfs	Natrixeralfs
Aridisols	Argids	Natragids
	Orthids	Salorthids
		Gypsiorthids
Entisols	Aquents	Sulfaquents
Inceptisols	Aquepts	Sulfaquepts
Mollisols	Aquolls	Natraquolls
	Borolls	Natriborolls
	Ustolls	Natrustolls
	Xerolls	Calcixerolls
		Natrixerolls

The depth of the natric horizon along the profile is not referred to in the US *Soil Taxonomy* or in the FAO/UNESCO Soil Map Legend. However, in the genetic soil classification system, the position of this horizon along the profile determines the subdivision of Solonetz soils:

- If the natric horizon is 0.8 cm - shallow Solonetz
- between 9-15 cm - medium Solonetz
- more than 16 cm - deep Solonetz

As a diagnostic feature, the natric horizon or B horizon is the most universal property among the main soil classification systems.

The physical and morphological properties of salt-affected soils are closely interrelated with their physico-chemical diagnostic properties, which jointly serve as crieteria in the soils' classification.

Chemical and Physico-chemical Soil Properties Related to the Classification of Salt-Affected Soils

<u>ESP, SAR values and related physico-chemical properties</u>

The sodium adsorption ratio (SAR) and the exchangeable sodium percentage (ESP) play significant roles. According to the US *Soil Taxonomy*, if SAR is $\geq$13 (or 15 percent or more saturation with exchangeable sodium) in some subhorizon within 40 cm of the upper boundary; or if there is more exchangeable magnesium plus sodium than calcium plus exchange acidity (at pH 8.2) in some subhorizon within 40 cm of the upper boundary if the SAR is $\geq$13 (or ESP $\geq$15) in some horizon within 2 m of the surface.

The 2 m depth in this definition seems too deep, because the diagnostic natric horizon must be much higher in the profile if it determines the soil type.

The diagnostics of Solonetz soils in the FAO/ UNESCO Soil Map Legend are very similar to those in the US *Soil Taxonomy*. In the genetic soil classification systems, the diagnostic characteristics besides the morphological features (see in (A)) are based mainly on the ESP values in the B horizon. The limits are very close to those in the US *Soil Taxonomy*.

It should be noted that the ESP values must be always considered together with the morphological features of the B horizon or illuvial horizon, because in some places typical morphology of natric horizon develops with comparatively low ESP and/or SAR values (e.g., ESP <7-10, in the Ukraine), while in other places no such natric horizon develops even in cases of high ESP value (ESP >20-25 in Sudan). The causes of such discrepancies can be different. Local differences in the definitions of appropriate diagnostic values of ESP in the sodic horizon need further study.

Nevertheless, the similarity of diagnostic ESP values in the different soil classification systems makes possible the correlation of sodic soils or alkali soils or solonetz soils or natric soils in the different soil classification systems.

<u>Salt content of the soil profile</u>

The salt content of soils (often called salinity) plays a significant role in all soil classification systems as diagnostic criteria for saline soils, solonchak soils, etc.

According to the US *Soil Taxonomy* the salic horizon is the main diagnostic feature for saline soils, which in this system belong mainly to the Aridisols. The diagnostics in FAO/UNESCO Soil Map Legend are similar for Solonchak soils. In the genetic soil classification system, too, analogously to the salic horizon, a saline horizon is the key diagnostic feature for Solonchak soils.

In the US *Soil Taxonomy*, the salic horizon is a horizon 15 cm or more thick that contains a secondary enrichment of salts more soluble in cold water than gypsum. It contains at least 2% salt.

In the FAO/UNESCO Soil Map Legend as well as in the genetic soil classification system, the limit values for salt content are somewhat lower; they do not restrict the salt accumulation to secondary enrichment. The genetic soil classification system also includes the gypsum content in the total salinity.

In spite of the fact that the convertibility of terms for the different saline soils is possible based on the main diagnostic features of the salt content of the salic horizon, certain discrepancies exist among the diagnostics of different soil classification systems in respect to salinity. The more important ones are as follows:

- While in the US and many other countries salinity commonly is determined by measuring the electric conductivity of soil paste or saturation extract (EC values), in the USSR, and many other countries where the genetic soil classification system is used, the chemical analysis of 1:5 or 1:1 aqueous extract is usual. Such differences make it difficult to convert the analytical values of classification units of different systems. Because of the different chemical composition of salts, the chemical analysis of aqueous extracts may produce salinity values which differ from those calculated from the EC values.
- The measurement of electric conductivity does not reveal the chemical composition of salts in the soil which is disclosed by the chemical analysis of an aqueous extract. EC determination does not show either whether the soluble salts, mainly sodium salts, are neutral, like sodium chloride, or are sodium salts capable of alkaline hydrolysis, like sodium carbonate or sodium bicarbonate.

The US *Soil Taxonomy* does not distinquish between saline soils dominated by neutral sodium salts and those dominated by sodium salts capable of alkaline hydrolysis. In the genetic soil classification system, soda Solonchak soils are separated from chloride and sulfate Solonchak soils. This approach should be considered because pedological properties, agronomical value, and the possibilities of reclamation of saline soils and reclamation strategies strictly depend on whether the soils have nearly neutral pH or strongly alkaline pH values.

The limit values for salinity should be different, depending on the chemistry of salt composition of the soils. Evidently if the salt carbonate dominates the salt composition, much lower concentrations result in adverse soil properties than if sodium chloride or sodium sulfate is dominant. Another condition exists when the presence of Mg salts dominate. This conditions is the easiest to reclaim.

A Few More Questions to be Answered

In spite of the fairly good correlation of the properties of salt-affected soils among the different classification systems, some problems need study, discussion, and agreement. From those only a few important ones are mentioned below.

(1) If we accept the definition of salt-affected soils described in the first part of this paper, gypsic soils, calcic soils, sulfatic and sulfidic, including acid sulfate, soils should be included in the system. In this case, according to the US *Soil Taxonomy*, the soil classification related properties of salt-affected soils can include calcic horizons and k horizons; partly gypsic, petrogypsic, and sometimes petrocalcic horizons; and even sulfuric horizons. (Also see Tables 1 and 2).

(2) In the practice of irrigation, the RSC (Residual Sodium Carbonate content in meq/L) can be a diagnostic feature for irrigation water quality. If the highly alkaline pH value of the soil and/or the free sodium carbonate plus biocarbonate content of the different chemical types of soil salinity, a more advanced classification could be elaborated which meets the practical requirements of irrigation and soil reclamation.

(3) The possibility of the occurrence of a salic and/or natric horizon in Vertisols and even Histosols should be studied.

Literature Cited

The Canadian System of Soil Classification. 1978. Canada Soil Survey Committee, Research Branch, Canada Department of Agriculture. Publication 1646.

FAO/UNESCO Soil Map of the World. 1988. Revised Legend. World Soil Resource Report 60, FAO, Rome.

Keys to Soil Taxonomy. 1985. Technical Monograph No. 6, Agronomy Department, Cornell University, Ithaca, New York.

Kovda, V.A. and B.G. Rozanov. 1988. Pochvovedenie (Soil Science) I-II. Vyshaya Shkola, Moscow.

Soils, an Australian viewpoint. 1983. CSIRO/Academic Press, Melbourne.

Szabolcs, I. 1989. Salt-affected soils. CRC Press Inc., Boca Raton, Florida.

Clayey Soils of Northern Canada and the Cordillera*

C. Tarnocai[1], G.F. Mills[2], H. Veldhuis[3], H. Luttmerding[4], and A. Green[5]

Abstract

Clayey soils which developed on glaciolacustrine, marine, and glacial till materials are found dispersed throughout northern Canada and the Cordillera. Permafrost occurs in all soils in the arctic, in many soils in the subarctic, and in some, usually poorly drained, soils in the boreal. Most of the soils containing permafrost have strongly cryoturbated horizons. In the boreal and the southern part of the Cordillera, well and imperfectly drained clayey soils usually have either a brownish B horizon (Bm) or a B horizon with significant clay accumulation (Bt). In the extreme southern part of the Cordillera, the clayey soils often have a moderately thick, organic-rich surface horizon (Ah). When located in poorly drained, depressional areas, they have gleyed horizons.

Arctic and subarctic clayey soils show no evidence of slickensides; cracking, if present, is due to thermal or frost cracking, rather than desiccation. Slickensides have been observed in some of the boreal (Manitoba) and Cordilleran (southern interior of British Columbia) clayey soils. Some of these boreal and Cordilleran soils also show evidence of cracking, but without significant grumic or churning characteristics.

According to present criteria, clayey soils in northern Canada and the Cordillera can not be classified in any of the Vertisol suborders of the American *Soil Taxonomy*. Vertic properties, if present, are not strong enough to warrant the soils being classified in the proposed Cryert suborder.

Introduction

The study area covers northern Canada, including the Arctic, Subarctic, and North Boreal Ecoclimatic Provinces, as well as the Cordilleran Ecoclimatic Provinces (Figure 1). This large area lies north and west of the Grassland and South Boreal Ecoclimatic Provinces (Figure 1), the subject of the presentation by Acton et al. (these proceedings).

Soils having clayey textures occupy small, but significant, areas in this large region. The most common materials on which these clayey soils have developed are glaciolacustrine and marine deposits, clayey glacial till, and residual materials derived mainly from shales.

The objective of this study is to describe the distribution and properties of clayey soils and their development as influenced by the range of climatic conditions and associated vegetation across the study area. The effects of cryogenic processes on the morphology of northern soils affected by permafrost is contrasted to the effect of shrink-swell processes on the morphology of southern clayey soils not affected by permafrost. The criteria used to classify permafrost-affected soils are compared with those used in the classification of Vertisols.

Physical Environment

Physiographic Regions

Physiographically and geologically, Canada is composed of two large and distinctly different areas: a core of old, massive, Precambrian crystalline rocks forming the Canadian Shield, and a surrounding younger, mainly sedimentary rock area forming the Borderlands (Figure 2). Further subdivisions of these two major physiographic units in northern and western Canada are shown in Figure 2, with the description derived mainly from the work of Bostock (1970) and Prest (1970). Clayey soils occurring throughout these areas have developed under a diverse range of climate and vegetation conditions (Figure 3).

A major part of Canada has been subjected to repeated glaciation during the Pleistocene ep-

*Land Resource Research Centre Contribution No.: 90-33.

[1]Land Resource Research Centre, Research Branch, Agriculture Canada, K.W. Neatby Bldg., Ottawa, Canada, K1A OC6.

[2]Manitoba Soil Survey, Room 362, Ellis Bldg., University of Manitoba, Winnipeg, Canada, R3T 2N2.

[3]Canada Soil Survey, Agriculture Canada, Room 362, Ellis Bldg., University of Manitoba, Winnipeg, Canada, R3T 2N2.

[4]British Columbia Ministry of the Environment, Policy and Planning Branch, 777 Broughton Street, Victoria, Canada, V8V 1X5.

[5]Canada Soil Survey, Agriculture Canada, 6660 N.W. Marine Drive, Vancouver, Canada, V6T 1X2.

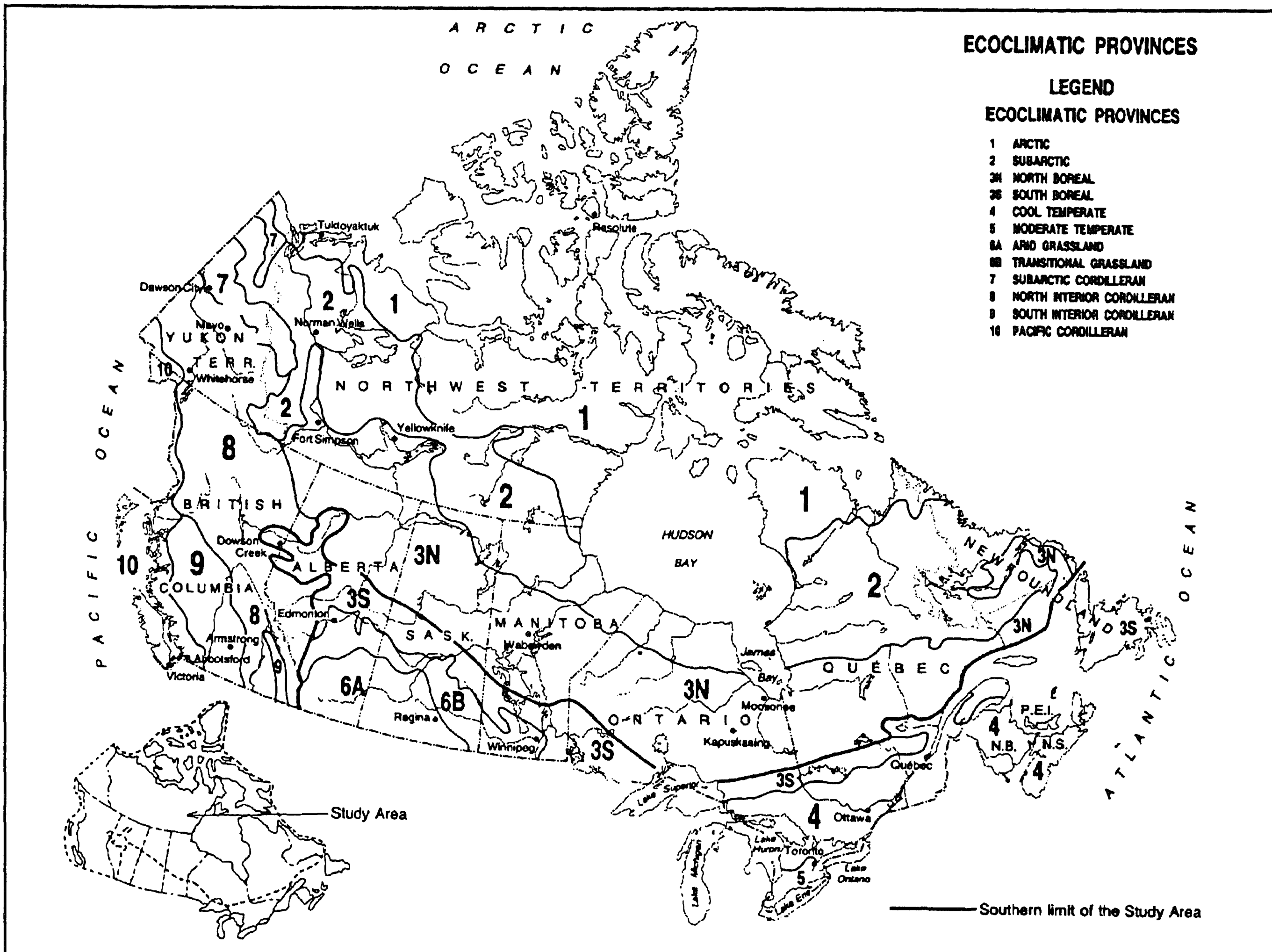


Figure 1. Ecoclimatic provinces of Canada (Ecoregion Working Group 1989).

och. As a result, the Canadian Shield is characterized by a thin cover of stony, sandy till. Fine-textured (silty clay and clay) glaciolacustrine deposits and gravelly or sandy beach materials have been deposited in the basins of former glacial lakes. These clayey glaciolacustrine deposits are common in north-central Manitoba and parts of northwestern Ontario, but sporadic or non-existent in the rest of the Canadian Shield.

The Hudson Region is a low-lying, peatland-dominated area on the south and west coasts of Hudson Bay. A large part of this region was inundated by the sea during the early Holocene epoch and clayey deposits are now found in the northern Ontario portion of this area.

The Interior Plains Region is composed of a series of plains, hills, and plateaus which rise gradually from the arctic coast southward. The area is covered mainly by clayey or loamy till. During the retreat of the glacial ice, a series of glacial lakes occupied most of the upper and central Mackenzie River Valley and the present Great Slave and Great Bear Lakes. Clayey and sandy deposits are common in the basins of these former glacial lakes.

The Arctic Lowlands Region consists of relatively gently rolling terrain. The loamy till cover is generally thin and bedrock outcrops are common. Clayey deposits usually are found only in areas along the coast which were inundated after the retreat of the glacial ice.

The Innuitian Region has a varied topography and is more rugged than the surrounding physiographic regions. The terrain is composed largely of bedrock associated with a veneer of sandy and loamy material. Along the coast the deposits are generally clays of marine origin.

The Arctic Coastal Plains Region is a strip 50 to 100 km wide along the shores of the Arctic Ocean from Meighen Island to Alaska. Some clayey marine deposits are found in the present-day Mackenzie River delta area and on Herschel

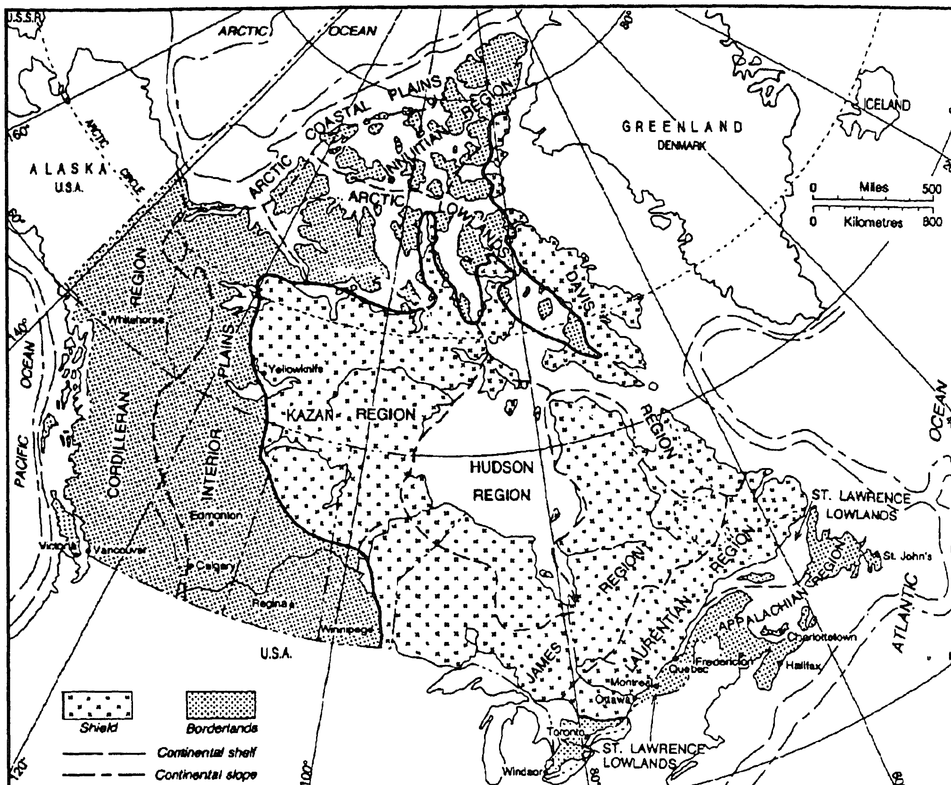


Figure 2. Physiographic regions of Canada (Bostock 1970).

Island. The eastern part of the Yukon coast is covered by clayey till; the central and western parts are characterized by clayey colluvial deposits.

The Cordilleran Region is divided longitudinally into three belts, the Eastern, Interior, and Western Systems, each with its own characteristic geology and physiography. Glacial lakes developed in valleys temporarily blocked by glacial ice. The deposits left behind in the larger basins often are composed of clayey materials. The largest of these occur in the central interior of British Columbia, the Whitehorse area of the southern Yukon, and the Old Crow area of the northern Yukon. Clayey materials of marine origin occur along the southern coastal areas of Vancouver Island and in the lower Fraser Valley of British Columbia.

Ecoclimatic Provinces

The ecoclimatic provinces of Canada were mapped and described by the Ecoregion Working Group (1989) and their distribution is shown in Figure 1. According to this work, the Arctic Ecoclimatic Province encompasses all those areas lying north of the arctic tree line, which reaches its most northerly limit in the Mackenzie River delta (approximately Lat. 69°) and its most southerly limit along the shores of Hudson Bay (approximately Lat. 57°). The Arctic Ecoclimatic Province occupies approximately 40% of the area of Canada and includes the northern continental area and all of the arctic islands.

The Subarctic Ecoclimatic Province encompasses the area south of the arctic tree line that has open coniferous forest vegetation. The North Boreal Ecoclimatic Province encompasses those areas south of the Subarctic Ecoclimatic Province dominated by closed canopy coniferous forest vegetation. The Cordilleran Ecoclimatic Provinces, which include the Subarctic Cordilleran, North Interior Cordilleran, South Interior Cordilleran and Pacific Cordilleran Provinces, encompass the mountains and plateaus of western and northwestern Canada. Climate-vegetation-soil relations throughout the Cordillera are governed by vertical as well as latitudinal zonation.

Atmospheric Climate

Climatic data for some selected stations in the study area are presented in Figure 4. The climate of the Arctic Ecoclimatic Province is characterized by long, cold winters and short, cool summers with low annual precipitation. Approximately 40 to 60% of the total precipitation occurs as snow. The winter is characterized by little daylight, the summer by long periods of daylight.

The climates of the Subarctic and North Boreal Ecoclimatic Provinces are characterized by long, cold winters and short, cool summers with moderate precipitation. The summers become warmer as one proceeds southward.

The climates of the Cordilleran Ecoclimatic Provinces vary greatly from south to north and with elevation. In the north the climate is similar to that described for the Subarctic Ecoclimatic Province. In the south, however, at low elevations the climate is temperate with warm, dry summers and cool winters; at higher elevations the climate becomes increasingly severe

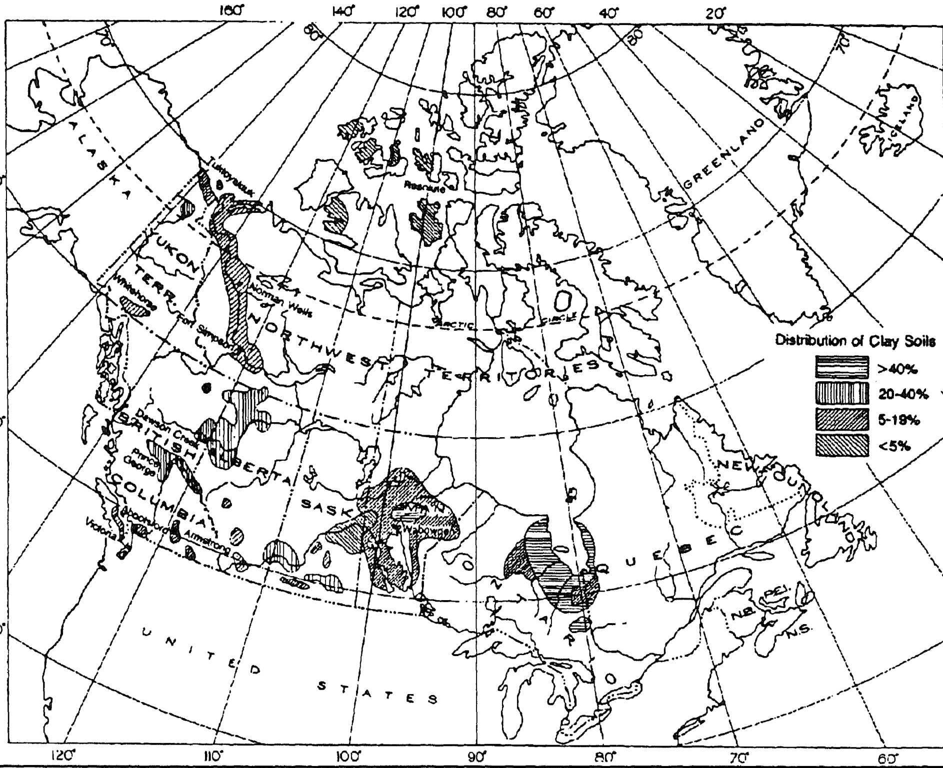


Figure 3. Distribution of clayey soils in the study area and the location of climatic stations.

and is similar to that of mountainous areas to the north. Along the Pacific Coast the climate is oceanic with mild winters, cool summers and high precipitation.

In Figure 4 both precipitation and Thornthwaite potential evapotranspiration are plotted for ten climatic stations. These graphs provide a general picture of the atmospheric moisture balance of the study area. In the Pacific Cordilleran (Victoria and Abbotsford stations) and South Interior Cordilleran (Armstrong) Ecoclimatic Provinces, a moisture deficiency occurs because of low precipitation during the summer. In the North Interior Cordilleran (Dawson Creek, Whitehorse, and Mayo), North Boreal (Wabowden, Fort Simpson, Kapuskasing, and Moosonee), Subarctic (Norman Wells), Subarctic Cordilleran (Dawson City) and Arctic (Tuktoyaktuk and Resolute) Ecoclimatic Provinces, most precipitation occurs during the summer months. The increased summer precipitation at the stations given for these areas, however, is not sufficient to compensate for evapotranspiration, and thus moisture deficiencies also occur. In the Ontario portion of the North Boreal Ecoclimatic Province, however, precipitation is significantly higher than in other portions of this ecoclimatic province, with the result that little moisture deficit occurs.

Throughout northern Canada and the Cordillera, soils with permafrost never completely dry out at depth, because of the gradual release of moisture to the soil as the seasonal frost melts. Soil moisture deficiency is therefore minimal in soils containing permafrost, even in areas that have an atmospheric moisture deficiency.

Soil Climate

The study area falls into four broad soil climatic regions: the Arctic, Subarctic, Cryoboreal, and Boreal (Clayton et al., 1977; Tarnocai, 1978), a modified version of which is shown in Figure 5. The Arctic Soil Climate is characterized by extremely cold soil temperatures. The MAST (mean annual soil temperature at 50 cm depth) is <-7°C and the MSST (mean summer soil temperature at 50 cm depth) is between 0°C and <3°C. The moisture regime of these soils ranges from aquic to humid.

The Subarctic Soil Climate is characterized by very cold soil temperatures. The MAST ranges from -7°C to <2°C, the MSST from 3°C to <6°C. The moisture regime of these soils ranges from aquic to humid.

The Cryoboreal Soil Climate is characterized by cold to moderately cold soil temperatures. The MAST ranges from 2°C to <5.5°C, the MSST from 6°C to <12°C. The moisture regimes of these soils are humid to subhumid in northern British Columbia and Alberta, humid to aquic in northern Saskatchewan and Manitoba, and subaquic to perhumid in northern Ontario.

The Boreal Soil Climate occurs only in the southern interior of British Columbia and is characterized by moderately cold soil temperatures. The MAST ranges from 5.5°C to <8°C, the MSST from 12°C to <18°C. The moisture regime of these soils ranges from subhumid to semiarid.

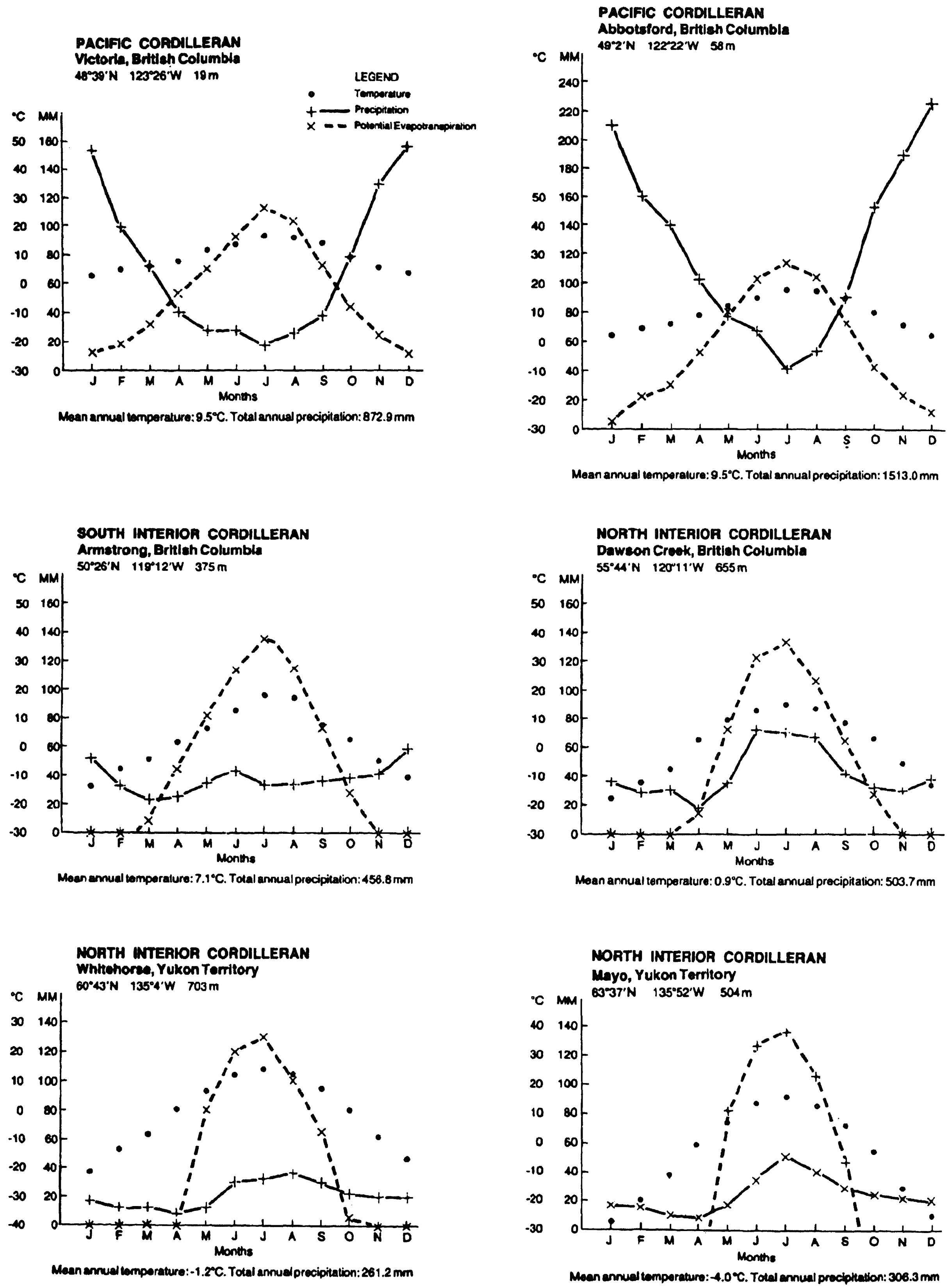


Figure 4. Mean monthly temperature and precipitation and Thornthwaite potential evapotranspiration for climatic stations in areas of major clay concentrations (Atmospheric Environment Service 1982, a and b). These climatic stations and the associated ecoclimatic provinces are as follows: Victoria, B.C. and Abbotsford, B.C. - Pacific Cordilleran; Armstrong, B.C. - South Interior Cordilleran; Dawson Creek, B.C. and Whitehorse, Y.T. - North Interior Cordilleran; Wabowden, Man. and Fort Simpson, N.W.T. - North Boreal; Norman Wells, N.W.T. - Subarctic; Tuktoyaktuk, N.W.T. and Resolute, N.W.T. - Arctic.

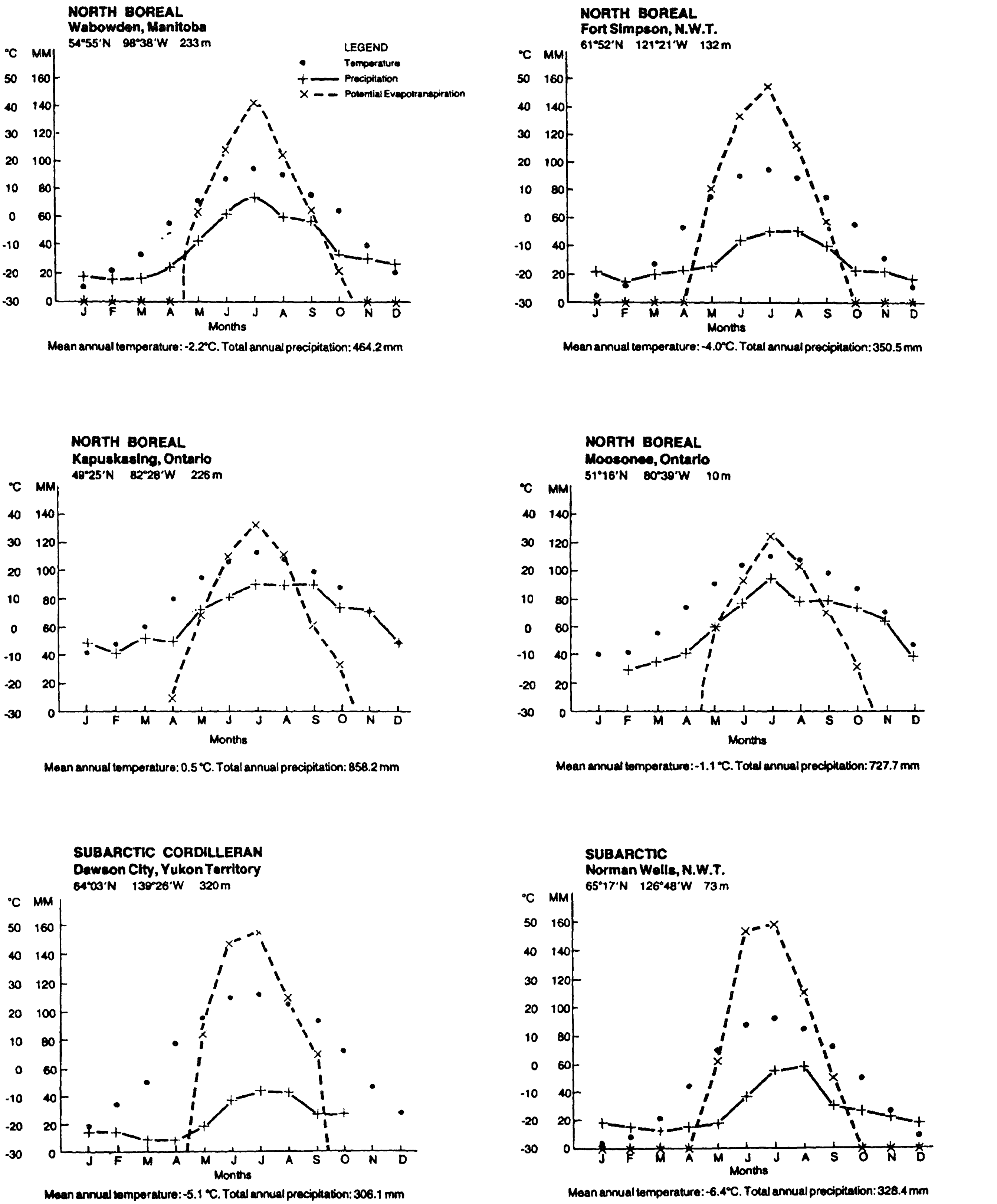

NORTH BOREAL
Wabowden, Manitoba
54°55'N 98°38'W 233 m
LEGEND
Temperature
Precipitation
Potential Evapotranspiration
°C MM
Months
J F M A M J J A S O N D
Mean annual temperature: -2.2°C. Total annual precipitation: 464.2 mm
NORTH BOREAL
Fort Simpson, N.W.T.
61°52'N 121°21'W 132 m
Mean annual temperature: -4.0°C. Total annual precipitation: 350.5 mm
NORTH BOREAL
Kapuskasing, Ontario
49°25'N 82°28'W 226 m
Mean annual temperature: 0.5 °C. Total annual precipitation: 858.2 mm
NORTH BOREAL
Moosonee, Ontario
51°16'N 80°39'W 10 m
Mean annual temperature: -1.1 °C. Total annual precipitation: 727.7 mm
SUBARCTIC CORDILLERAN
Dawson City, Yukon Territory
64°03'N 139°26'W 320 m
Mean annual temperature: -5.1 °C. Total annual precipitation: 306.1 mm
SUBARCTIC
Norman Wells, N.W.T.
65°17'N 126°48'W 73 m
Mean annual temperature: -6.4°C. Total annual precipitation: 328.4 mm

Figure 4.

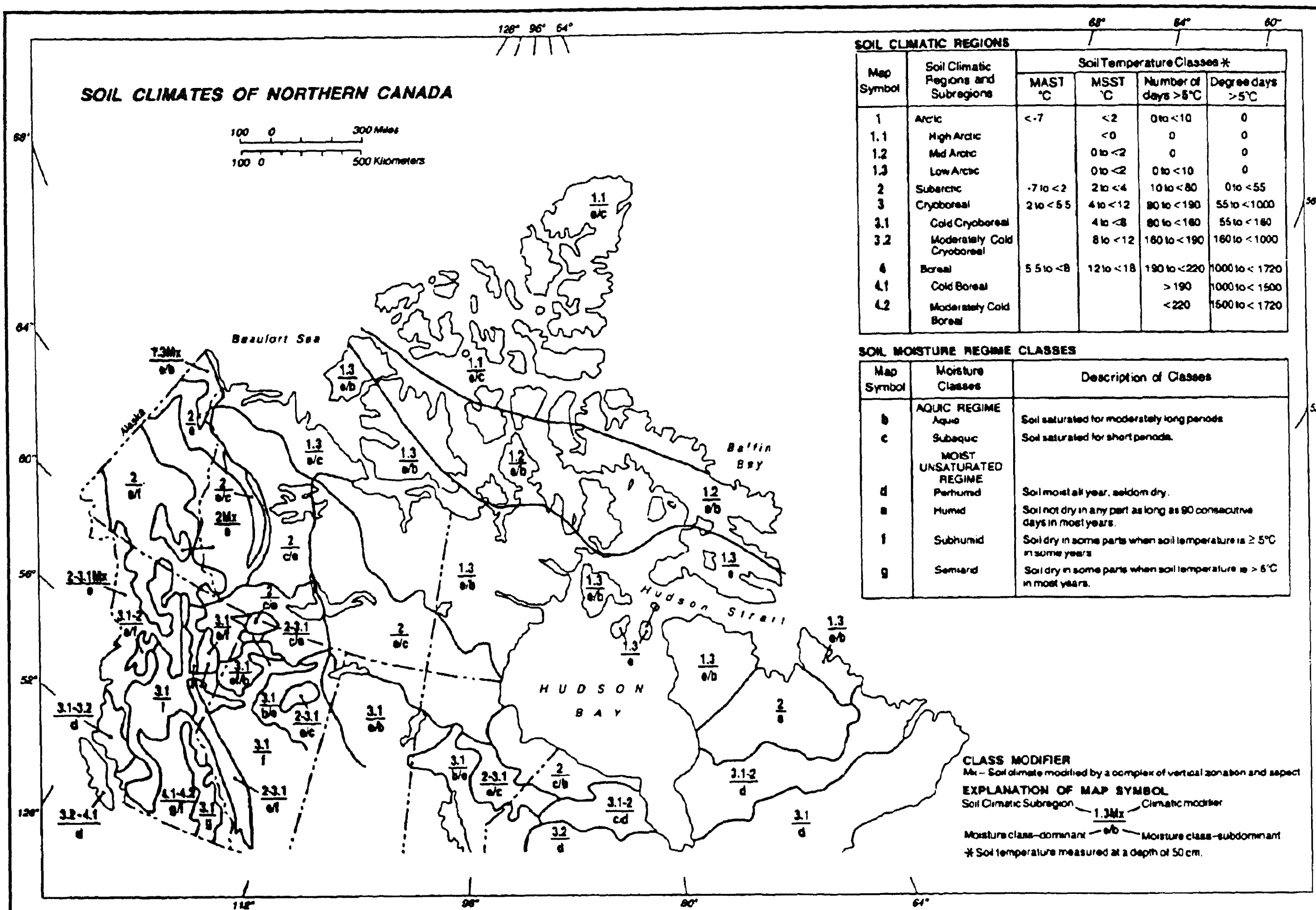


Figure 5. Soil climates of the study area (Tarnocai; 1978, Clayton et al., 1977).

Permafrost

Permafrost is a thermal condition in soil or rock in which temperatures below 0°C persist over at least two consecutive winters and the intervening summer (Harris et al., 1988; Brown and Kupsch, 1974). Approximately one-half of Canada lies in the permafrost region (Figure 6), which is further divided into two zones - discontinuous in the south and continuous in the north (Brown 1967).

In the southern part of the discontinuous zone, permafrost occurs as scattered islands a few square metres to several hectares in size. The permafrost is confined to peatlands, wet clayey soils, and north-facing slopes. In the northern part of this zone, the permafrost becomes increasingly widespread and occurs in a great variety of terrain types. Finally, in the continuous zone, permafrost exists in all terrain types under all moisture conditions. The only exception is in newly deposited sediments where the climate has not yet imposed its influence on the soil thermal regime.

Soils of the Study Area

All areas of the Arctic Ecoclimatic Province are covered by Cryosolic soils (Bentley ed., 1978; Tarnocai, 1978). Most of these Cryosols (Agriculture Canada Expert Committee on Soil Survey, 1987), classified as Pergelic Inceptisols (Soil Survey Staff, 1975) in the American soil classification system (see Table 1 for the correlation between the two soil classifications), are strongly cryoturbated and are associated with permafrost, patterned ground, and tundra vegetation.

Cryosolic soils dominate the Subarctic Ecoclimatic Province, although Brunisolic soils (Inceptisols, Soil Survey Staff, 1975) are also present in the southern part of this province. Patterned ground features are common, especially polygons, earth hummocks, and circles.

The Boreal Ecoclimatic Province is dominated by Brunisolic and Luvisolic soils (Inceptisols, Boralfs and Udalfs, Soil Survey Staff, 1975). Organic soils (Histosols, Soil Survey Staff, 1975) and some Cryosols are found on peat materials.

The Cordilleran Ecoclimatic Provinces are dominated by Luvisols, Podzols (Spodosols, Soil Survey Staff, 1975) and Brunisols, with smaller amounts of Chernozemic (Mollisols, Soil Survey Staff, 1975) and Cryosolic soils.

Properties of Clayey Soils

Some Properties of Clayey Soils in the Yukon and Northwest Territories

Most of the Northwest Territories lies within the Arctic and Subarctic Ecoclimatic Provinces, although a small portion lies within the North Boreal Ecoclimatic Province. The Yukon has a more complex ecoclimate, dominated by the North Interior Cordilleran Ecoclimatic Province. Also present in significant areas are the Subarctic Cordilleran and Subarctic Ecoclimatic Provinces. Only the extreme northern portion of the Yukon lies in the Arctic Ecoclimatic Province (Figure 1). Most of the information presented in this section refers to the clay soils occurring in the North Boreal, Subarctic, and Arctic Ecoclimatic Provinces.

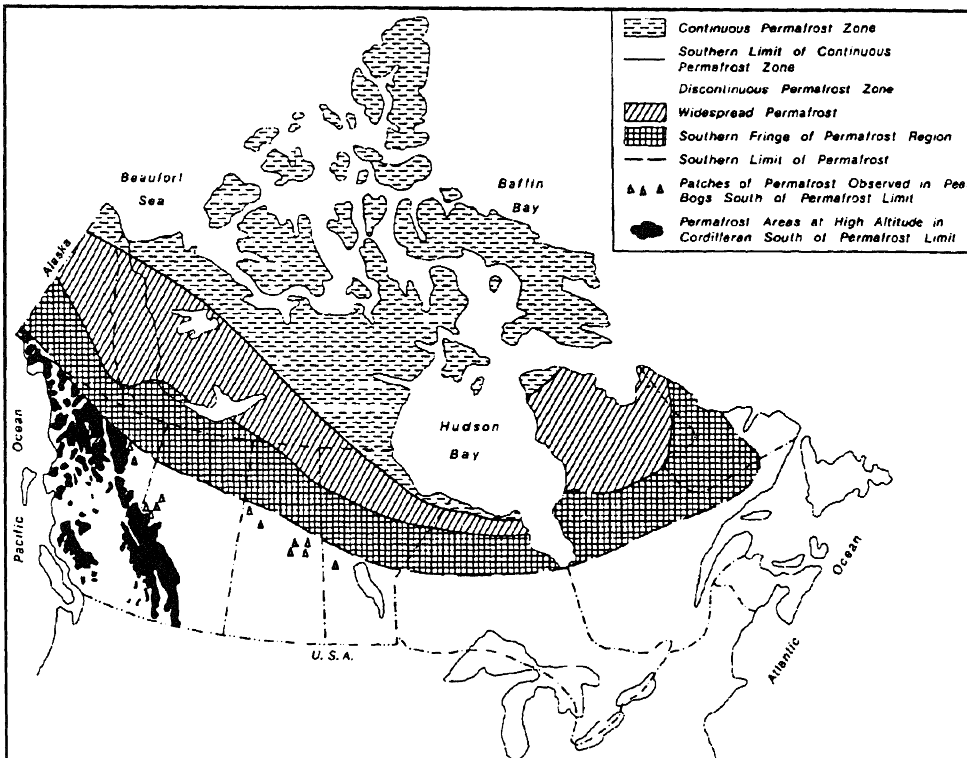


Figure 6. Permafrost regions of Canada (Harris et al., 1988; Brown, 1967).

Table 1. Correlation between the Canadian and American soil classification systems.*

Canadian Soil Classification[1]	American Soil Classification[2]
Brunisol	Inceptisol
Orthic Eutric Brunisol	Cryochrept, Eutrochrept
Orthic Dystric Brunisol	Cryochrept, Dystrochrept
Chrnozem	Boroll
Orthic Dark Gray Chrnozem	Boralfic Boroll, Argic Boroll
Eluviated Dark Brown Chernozem	Typic Boroll
Cryosol	Pergelic subgroups
Orthic Turbic Cryosol	Pergelic Ruptic Cryochrept
Brunisolic Turbic Cryosol	Pergelic Ruptic Cryochrept
Orthic Static Cryosol	Pergelic subgroups
Gleysol	Aqu-suborders
Humic Luvic Gleysol	Aquoll, Humaquept
Luvisol	Boralf & Udalf
Orthic Gray Luvisol	Cryoboralf
Solonetzic Gray Luvisol	Cryoboralf
Organic	Histosol
Podzol	Spodosol
Grumic Families of various orders	Vertisols

*Only those soils included in this paper are correlated.
[1]Agriculture Canada Expert Committee on Soil Survey, 1987.
[2]Soil Survey Staff, 1975.

Morphology

The clayey soils of the Arctic and Subarctic Ecoclimatic Provinces are classified as Cryosols and are associated with permafrost. The permafrost table usually occurs within 80 cm of the surface. Earth hummocks, a type of patterned ground, provide the microrelief (Figure 7), ranging in height from 40 to 60 cm and in diameter from 80 to 160 cm (Tarnocai and Zoltai, 1978). In the High Arctic Ecoclimatic Subprovince clayey soils are very often severely cracked and heaved (Figure 8).

The internal morphology of these soils is dominated by cryogenic features such as displaced soil horizons, mixed soil materials, and organic smears** resulting from the movement of soil materials by cryoturbation. An example of a cryoturbated clayey soil associated with earth hummocks is shown in Figure 9.

The granular structure of the surface mineral horizons of these clayey Cryosolic soils is the result of repeated freezing and thawing cycles and the associated movement of soil materials. The granular horizons grade into massive, structureless subsurface horizons (Table 2) overlying a perennially frozen horizon in which ice lenses, vein ice, and pure ice layers commonly occur. Organic materials occur in all subsurface horizons, including the frozen hori-

**Organic smears result from movement of soil materials by cryoturbation. When organic materials are moved they leave a trail of thin, darker organic material smeared on the lighter mineral material.

zon. This organic material has been moved downward by cryoturbation and forms distinct bodies, layers, and smears.

In the southern part of the Yukon (North Boreal Ecoclimatic Province), the well to imperfectly drained clayey soils are Brunisols. These soils have B horizons with subangular blocky structure. The generally level or, in some cases, slightly hummocky microrelief is masked by a layer of forest humus (L, F, and H horizons). In poorly drained areas, the clayey soils generally have a peaty surface layer. Permafrost is common in these poorly drained soils, as a result of the insulating properties of this peat layer.

Physical Properties

The clay contents of the B horizons of some clayey soils are presented in Table 3 and range from 24 to 60%. Very slight clay accumulation sometimes occurs in the B horizons of clayey Brunisolic soils, but no such accumulation occurs in any of the horizons of the clayey Cryosolic soils.

The clay mineralogy for Soil 3 (Table 2) is given by Pettapiece et al. (1978) and is dominated by smectite, vermiculite, and mica. Traces of chlorite and kaolinite are also present.

Soil Moisture and Ice Content

The near surface soil materials of clayey Cryosolic soils have volumetric moisture contents of approximately 30%. The moisture contents begin to increase rapidly close to the permafrost table. The increases are primarily the result of water being released from the ice-rich subsoil. The volumetric moisture (ice) content of the frozen soil materials is generally 30 to 50%, although in ice-rich layers it can be much higher, reaching 100% in pure ice layers (Figure 10).

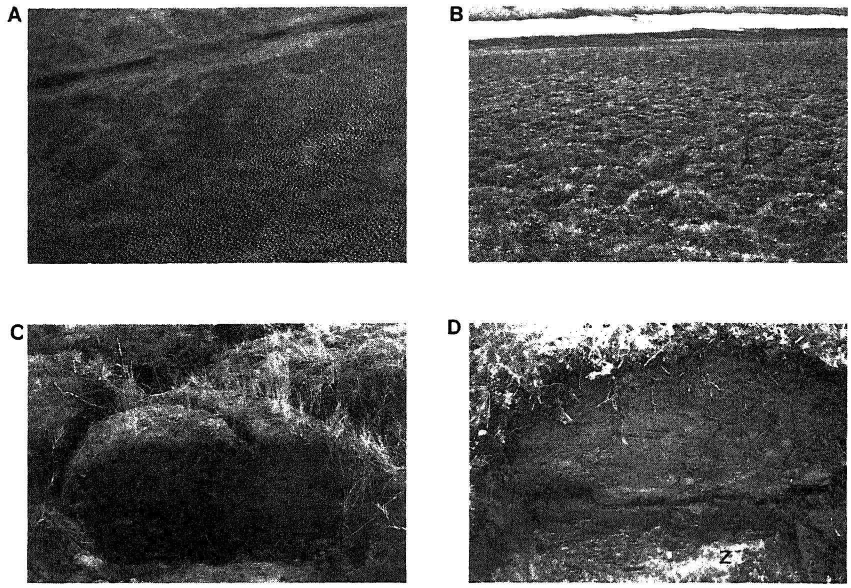


Figure 7. Earth hummocks associated with clayey soils in the Canadian arctic. Earth hummocks viewed (A) from the air and (B) from the ground. Cross sections of earth hummocks (C) with a frost crack and (D) with an icerich perennially-frozen soil horizon (z).

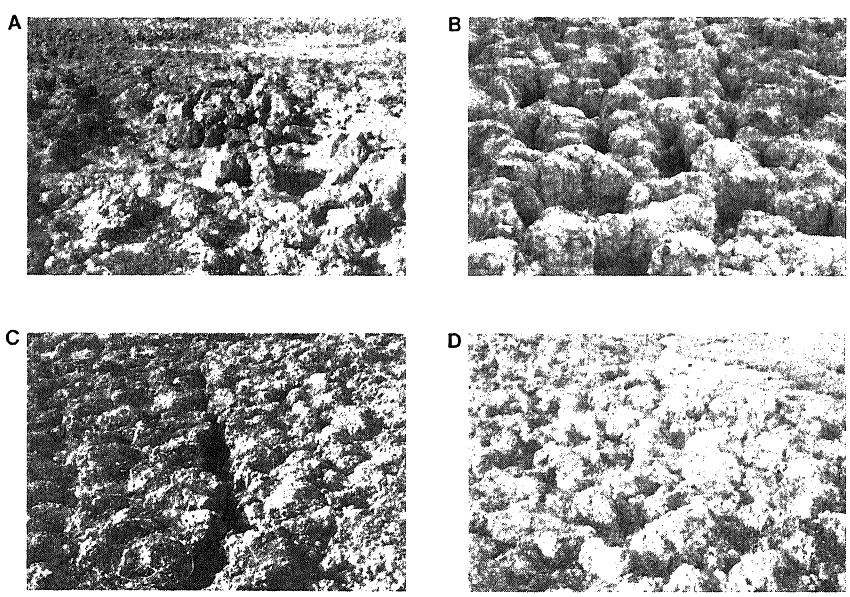


Figure 8. Severely frost-cracked and heaved lacustrine clayey soils in the Lake Hazen area, northern Ellesmere Island (Lat. 81° 49'N, Long. 71° 18'W). Photograph C shows a frost crack which is about 10 cm wide.

Active Layer

The active layer is that layer of the soil, lying above the permafrost, that is subject to annual thawing and refreezing. Its thickness is con-

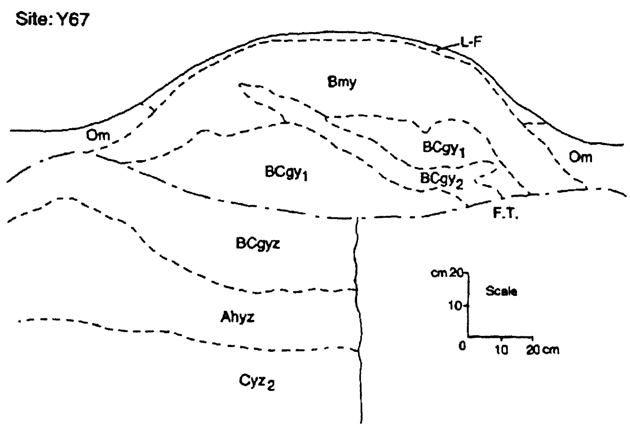


Figure 9. Cross section of an earth hummock developed on clayey materials, showing the cryoturbated soil horizons (Soil 5, Table 3). Location: 68° 46'N 134° 03'W (N.W.T.).

trolled by the soil texture and moisture, the surface organic cover, the vegetation cover, and the latitude. The relationships between the active layer depth and these factors are shown in Figure 11. The data presented in this figure indicate that the thickness of the active layer in well to imperfectly drained clayey soils ranges from 27 to 38 cm in the High Arctic Ecoclimatic Subprovince, from 35 to 56 cm in the MidArctic subprovince, from 34 to 62 cm in the Low Arctic subprovince, and from 70 to 98 cm in the Subarctic Ecoclimatic Province (Tarnocai, 1978). The thickness of the active layer in poorly drained clayey soils and soils with peaty surface layers is 10 to 20 cm less than that of well to imperfectly drained clayey soils in the Arctic Ecoclimatic Province and 30 to 50 cm less in the Subarctic Ecoclimatic Province (Tarnocai, 1978).

Frost Cracking

Frost cracking is the fracturing of soil material by thermal contraction in subfreezing temperatures (Washburn, 1979) (Figure 12). Ice is the critical material in frozen soils. Pure ice has a coefficient of linear contraction of about $45x10^{-6}$ °C-1 at -40°C. The factors affecting cracking are the salt content, the highly variable distribution of ice in the frozen soil, and the insulating effect of the snow.

Frost cracks can develop in both sandy and clayey soil materials as well as in organic materials. Fine textured, moisture-rich soils are probably the soils which are most susceptible to frost cracking. In permafrost soils, the initial frost cracks start at the surface and can extend to a depth of 4 m (Mackay 1972, 1974).

Cracks are primarily imprinted in the permafrost layer, rather than in the active layer. Thus, once the frost cracks are initiated, they recur in the same place year after year. The

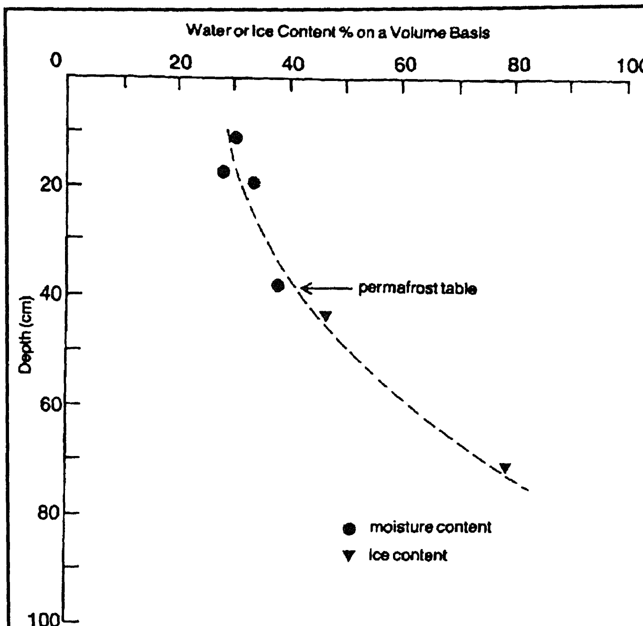


Figure 10. Moisture and ice content of a clayey hummocky soil from the Canadian North. The permafrost table is located at the 60 cm depth. Location: 68° 46'N 134° 03'W (N.W.T.).

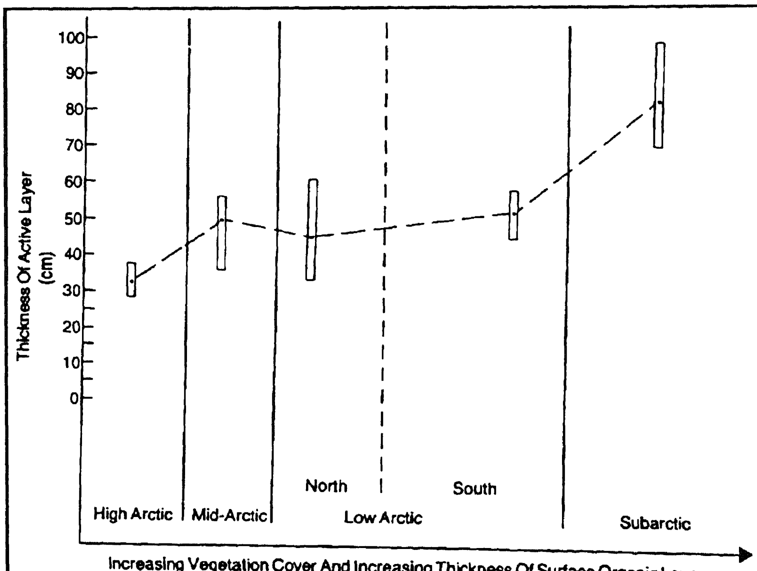


Figure 11. Thickness of the active layer in various clayey soils in the Arctic and Subarctic Ecoclimatic Provinces (Tarnocai 1978). The range of active layer thicknesses is indicated by a bar and the mean active layer thicknesses are indicated by a broken line (- - -).

Table 2. Morphological description of some northern clayey soils.

Horizon	Depth (cm)	Colour (moist)	Texture*	Structure+	Special Features
	1. Orthic Eutric Brunisol 60° 50'N, 135° 20'W				
Ah	3-0	10YR 2/1	—	—	
Ba1	0-13	10YR 4/3	SiCL	sbk	
Ba2	13-28	10YR 5/4	SiCL	sbk	
Ba3	28-39	10YR 5/3	SiC	sbk	
Ck	39+	7.5Y 6/2	SiC	pr	
	2. Orthic Eutric Brunisol 60° 40'N, 119° 04'W				
L, H	8-0	10YR 3/2	—	—	
Bm	0-14	7.5YR 5/4	C	sbk	
BC	14-55	10YR 6/3	SiC	sbk	
C	55-100	10YR 7/1	SiC	sbk	
	3. Orthic Turbic Cryosol 63° 37'N, 123° 39'W				
Of1	24-9	7.5YR 3/2.5	—	—	
Of2	9-4	10YR 2/1.5	—	—	
Om	4-0	5YR 2.5/1.5	—	—	
Bmy	0-22	5YR 3.5/1	SiC	gr	
BCy	22-47	10YR 3/1	SiC	mss	strongly cryoturbated
Ahyz	47-59	10YR 3/1	SiC	mss	strongly cryoturbated
Cyz	59+	10YR 4/1	SiC	mss	strongly cryoturbated
	4. Orthic Turbic Cryosol 68° 46'N, 134° 03'W				
L, H	1-0	10YR 3/2	—	—	
Bmy	0-20	10YR 5/4	C	gr	strongly cryoturbated
BCgy	20-60	10YR 5/2	C	gr	strongly cryoturbated
Cyzl	60-88	5Y 5/1	CL	mss	strongly cryoturbated
Ahyz	88-100	10YR 3/1	CL	—	strongly cryoturbated
	5. Brunisolic Turbic Cryosol 68° 23'N, 133° 44'W				
Om	10-0	5YR 2.5/1	—	—	
Bm	0-20	10YR 4/3	C	gr	
BCy1	20-70	10YR 4/3	C	mss	strongly cryoturbated
BCy2	70-100	10YR 4/2	C	mss	strongly cryoturbated
Ohy	100-110	5YR 3/2	—	—	strongly cryoturbated
Cg	110-130	10YR 4/1	C	mss	
Cz	130+	10YR 4/1	C	mss	
	6. Orthic Turbic Cryosol 69° 35'N, 139° 05'W				
Ahky	0-10	10YR 2/2	SiC	sbk	
Bmky	0-17	5Y 5/2	SiC	gr	strongly cryoturbated
Cky1	0-28	10YR 3/2	SiC	sbk	strongly cryoturbated
Cky2	0-40	10YR 3/2	SiC	pr	strongly cryoturbated
Ckz	-	5Y 5/2	SiC	—	

*Texture: C-clay; CL-clay loam; SiC-silty clay; SiCL-silty clay loam
+Structure: sbk-subangular blocky; gr-granular; mss-massive; pr-prismatic

crack can occur as a single crack more or less linear in form, but most commonly, the cracks occur in a polygonal pattern (Figure 12C). The cracks are usually filled by snow and melt water or, in dry high arctic areas, by sand. As a result, frost cracks in permafrost soils are commonly occupied by either ice wedges (Figure 12D) or sand wedges. In the arctic regions, however, ice wedges are very common and sand wedges are relatively rare.

Cryoturbation

Most northern Canadian soils are subject to cryoturbation, a process which affects the entire soil system. As a result of cryoturbation, the soil surface is unstable and, internally, soil materials are mixed together and soil horizons disrupted and displaced. These properties are associated with Turbic Cryosols and with those Brunisols and Gleysols which occur in the northern part of the North Boreal Ecoclimatic Province or above the timberline in some mountainous areas.

Chemical Properties

The sola of clayey Cryosolic soils are acid to mildly alkaline, ranging in pH between 3.8 and 7.9 (Table 3). The sola of clayey Brunisolic soils are slightly acid to neutral (Soils 1 and 2, Table 3) with pH values somewhat lower than those of the parent material as a result of leaching.

The organic carbon content of the clayey Brunisolic soils decreases with depth. The clayey Cryosolic soils, on the other hand, have high levels of organic carbon in all mineral horizons as a result of cryoturbation. In the Cryosolic soils, an organic-rich mineral horizon commonly occurs in the lower part of the pedon, in the vicinity of the permafrost table.

The cation exchange capacity ranges between 10 and 20 meq/100g, with calcium and magnesium being the dominant exchangeable cations (Table 3). Although the electrical conductivity is generally low (0.1 to 0.3 mmhos/cm), in some cases, the parent material has a higher conductivity (Cz horizon of Soil 6, Table 3). In the Arctic Ecoclimatic Province, clayey soils very often develop a salt crust on the soil surface during the warmer summer days, especially in periods with no precipitation when moisture is drawn to the surface where it evaporates.

Present Use

Almost all of the clayey soils occurring in the Yukon and Northwest Territories are under native forest or tundra vegetation. The main problems associated with the use of these soils result from the high amounts of ice in the frozen subsoil. Removal of the vegetation and organic surface horizons causes the soil temperature to rise and the ice in the subsoil to melt. This results in pronounced subsidence and an uneven topography.

When these soils occur on sloping topography, they are subject to solifluction and erosion if the thermal balance is disturbed. Clayey soils with level topography are very often subject to thermokarst. This thermokarst process results from the melting of ice-rich material and leads to subsequent thaw settlement. All activities involving these high ice content clayey soils should be carried out in such a way that melting of the ice is minimized.

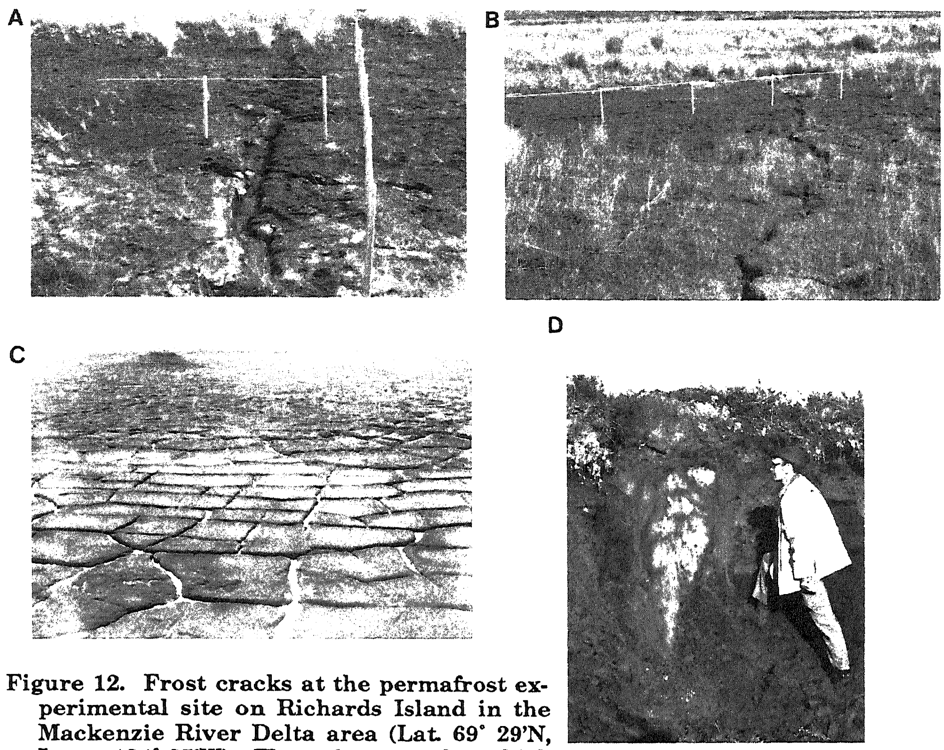


Figure 12. Frost cracks at the permafrost experimental site on Richards Island in the Mackenzie River Delta area (Lat. 69° 29'N, Long. 134° 35'W). These frost cracks, which are seven years old and about 15 to 30 cm wide, have developed in drained lake sediments (A and B). In these two photographs the vertical stakes with the horizontal rods are used to monitor the widening of the frost crack. Ice wedge polygons with wide (1 to 3 m) frost cracks (trenches), some filled with water (C). The ice wedges are beneath the trench. An ice wedge developed in a polygon trench (D). Photographs C and D were taken in the Mackenzie River Delta area.

Some Properties of Clayey Soils in the Northern Portions of Ontario, Manitoba, and Saskatchewan

Most of this area lies within the North Boreal Ecoclimatic Province. Only the extreme northern part of the area lies within the Subarctic Ecoclimatic Province (Figure 1). The information presented in this section refers only to the North Boreal Ecoclimatic Province, since this is where most of the clayey soils occur.

Morphology

Gray Luvisols have developed on clayey materials throughout the North Boreal Ecoclimatic Province (Figure 2), under the influence of forest vegetation. Their main characteristics are continuous light-coloured eluvial horizons and illuvial textural B horizons. These developed through leaching of the soluble decomposition products of forest litter and consequent downward translocation and immobilization of clays with other associated colloidal materials.

Increasingly severe climate towards the northern edge of the North Boreal Ecoclimatic Province (Figure 2) results in the persistence of permafrost below the 1 m control section in some Luvisols and the widespread occurrence of permafrost within 1 m in Static Cryosols and 2 m in Turbic Cryosols. The Cryosols are usually associated with varying degrees of cryoturbated horizons and patterned ground features on the soil surface.

Other clayey soils in these northern landscapes include poorly drained Gleysolic soils with thin surface veneers of peat. Very poorly drained lower slopes and depressions are characterized by Organic soils in which peat accumulation ranges from 40 cm to in excess of 1.6 m. These Organic soils are underlain by clay sediments.

The most southerly occurrence of permafrost is in organic materials; on progressing northward, permafrost occurs next in poorly drained clay soils and eventually is encountered in well drained clays near the northern boundary of the North Boreal Ecoclimatic Province. The occurrence of permafrost affects soil morphology in terms of cryoturbation, soil moisture regime, ice content and active layer depth.

Variation in degree of cryoturbation in well to imperfectly drained clay soils is summarized in Table 4. Soil 1 has developed continuous undisturbed horizons. Soil 2 is characterized by low microhummocky relief with continuous horizons (Figure 13a). The hummocky surface of this soil is a relict feature from a climatic period during which cryoturbation and earth hummock formation occurred. Soil 3 occurs in the Subarctic Ecoclimatic Province and is affected by permafrost and more pronounced cryoturbation (Figure 13b). This Cryosolic soil is weakly developed with irregular, broken and convoluted horizons and strongly mounded microrelief.

Physical and Chemical Properties

Clayey parent materials in northern Ontario, Manitoba, and Saskatchewan are generally moderately to strongly calcareous, although a

few areas of neutral, noncalcareous clay occur. The mineralogy of these materials is mixed; high shrink-swell clays such as smectite originate from the shale bedrock of the Interior Plains Region to the south and west. Dilution of smectite with illitic clay derived from Precambrian crystalline rock occurs in many of the clayey sediments in the Canadian Shield. Physical and chemical properties of three clay soils from northern Manitoba are presented in Table 5.

All soils have been leached to varying degrees (removal of soluble salts and $CaCO_3$ and reduced pH in the sola). The well developed Luvisols (Soils 1 and 2, Table 5) have deeper sola and distinct horizonation resulting from translocation of clay. The clayey Cryosol (Soil 3, Table 5) is leached to a shallow depth, with weakly developed horizons and no pronounced accumulation of translocated clay. Development of this soil is impeded by the permafrost table and the unstable soil surface resulting from cryoturbation.

Table 3. Some selected physical and chemical characteristics of some northern clayey soils.

Horizon	Depth (cm)	Total sand	Silt (%)	Clay (%)	pH	Cond. (mmhos/cm)	C	CEC (%)	Exchangeable Cations (meq) Ca	Mg	meq/100 g K	Na
	1. Orthic Eutric Brunisol 60° 50'N, 135° 20'W											
Ah	3-0	—	—	—	—	—	—	—	—	—	—	—
Bm1	0-13	10.0	56.7	33.2	5.4	0.1	2.3	—	13.9	12.5	0.8	0.1
Bm2	13-28	2.7	63.6	33.7	6.2	0.1	1.1	—	12.1	10.2	0.7	0.2
Bm3	28-39	1.2	58.0	40.7	6.4	0.1	1.0	—	14.0	12.7	0.5	0.3
Ck	39+	4.9	49.8	45.3	—	0.1	2.1	—	31.3	12.5	0.1	0.5
	2. Orthic Eutric Brunisol 60° 40'N, 119° 04'W											
L, H	8-0	—	—	—	6.3	0.4	52.3	88	71.9	7.6	0.6	0.2
Bm	0-14	24.6	34.8	40.6	6.9	0.3	1.9	24	20.7	2.6	0.4	_Tr*
BC	14-55	4.1	45.9	50.0	7.7	0.2	1.3	19	—	—	—	—
C	55-100	6.2	65.4	28.3	7.8	0.2	—	10	—	—	—	—
	3. Orthic Turbic Cryosol 63° 37'N, 123° 39'W											
Of1	24-9	—	—	—	3.7	—	41.9	—	20.9	9.0	1.4	—
Of2	9-4	—	—	—	5.9	—	37.2	—	88.6	25.5	0.7	—
Om	4-0	—	—	—	5.9	—	33.4	—	96.8	26.6	0.2	—
Bmy	0-22	9.0	49.7	41.3	6.7	0.10	2.3	—	17.8	7.6	0.2	—
BCy	22-47	7.3	52.7	39.9	6.7	0.10	8.4	—	42.8	12.3	0.2	—
Ahyz	47-59	9.5	56.0	43.5	6.7	0.17	10.4	—	46.6	11.5	0.3	—
Cyz	59+	8.4	54.5	37.2	6.8	0.15	8.5	—	43.6	11.3	0.5	—
	4. Orthic Turbic Cryosol 68° 46'N, 134° 03'W											
L, H	1-0	—	—	—	4.5	—	44.2	—	12.7	8.9	4.1	0.3
Bmy	0-20	18.6	36.2	45.2	3.8	0.10	2.5	22	0.4	1.9	0.3	Tr*
BCgy	20-60	19.3	36.4	44.3	4.0	0.05	1.5	20	1.3	2.1	0.2	0.1
Cyz1	60-88	27.0	37.7	35.2	4.4	0.28	4.7	22	3.5	3.8	0.6	0.1
Ahyz	88-100	23.4	39.9	36.8	4.4	0.32	12.9	33	6.2	5.3	0.5	0.1
Cyz2	100-125	22.2	40.0	37.8	4.3	0.50	—	21	7.5	3.3	0.7	0.4
	5. Brunisolic Turbic Cryosol 68° 23'N, 133° 44'W											
Om	10-0	—	—	—	—	—	—	—	—	—	—	—
Bm	0-20	3.0	49.0	48.0	3.8	—	2.0	27	0.1	0.2	0.3	0.1
BCy1	20-70	2.0	45.0	53.0	3.7	—	1.2	25	0.2	0.3	0.5	0.1
BCy2	70-100	3.0	47.0	50.0	3.7	—	2.3	27	0.6	0.5	0.6	0.1
Ohy	100-110	—	—	—	4.1	—	17.4	78	2.4	1.3	0.5	0.2
Cg	110-130	2.0	48.0	50.0	3.7	—	2.7	25	6.8	3.8	0.7	1.2
Cz	130+	—	—	—	—	—	—	—	—	—	—	—
	6. Orthic Turbic Cryosol 69° 35'N, 139° 05'W											
Ahky	0-10	5.4	43.3	51.3	7.4	0.50	4.5	—	—	—	—	—
Bmky	0-17	2.9	37.2	59.9	7.6	0.30	2.2	—	—	—	—	—
Cky1	0-28	4.0	49.2	46.8	7.6	0.20	2.6	—	—	—	—	—
Cky2	0-40	3.1	48.8	48.2	7.5	0.20	2.6	—	—	—	—	—
Ckz	-	3.8	51.2	45.0	7.5	0.20	2.9	—	—	—	—	—

*Tr - trace

Soil Moisture and Ice Content

Clayey Luvisolic soils occur under moisture regimes in which annual precipitation and potential evapotranspiration result in only short periods of soil moisture deficit (Figure 4, Wabowden). The moisture regime of clayey Cryosolic soils is influenced by the presence of permafrost and the slow release of moisture from the ice-rich subsoil as it thaws during the growing season. As a result, Cryosols may become dry to a very shallow depth for only short periods.

Active Layer

The depth of the active layer in well and imperfectly drained clayey Luvisols containing permafrost varies from 0.7 to 2 m in northern portions of the North Boreal Ecoclimatic Province. Although these soils do not contain permafrost in the control section in all years, they are the coldest Luvisols. In the Subarctic Ecoclimatic Province, the active layer in well and imperfectly drained clayey soils decreases to about 0.5 to 1.2 m, depending on organic surface material thickness and vegetative cover. Some part of the control section remains frozen throughout the year.

Cryoturbation

Cryoturbation is evident in Luvisols in which permafrost occurs only periodically within the control section and also in Luvisols no longer affected by permafrost. Subdued earth hummocks occur on the surface of these cold Luvisols, but horizon development is continuous and not disrupted or convoluted (Figure 13a). Such patterned ground is a relict feature from climatic periods when cryoturbation was more active. Clayey Cryosols characterized by permafrost within the control section have shallow active layers and a strongly mounded microhummocky surface pattern. Cryoturbation is sufficiently active to disturb and disrupt soil horizons, cause unstable soil surfaces, and permit only relatively weak horizon development (Figure 13b).

Vertic Properties

Clayey Luvisolic soils have clay content and mineralogy that have the potential to develop vertic properties. Vertic features have been observed in well and imperfectly drained Luvisolic clay soils, but the frequency and extent of their occurrence has not been the subject of any quantitative study. Cracks that are open to the surface seldom extend below the upper solum of Luvisolic soils in the North Boreal Ecoclimatic Province.

Slickensides have been observed in southern Luvisols in the study area, but appear to be uncommon in more northerly clayey soils, particularly those affected by permafrost and cryoturbation (Mills et al., these proceedings). This can be attributed to shorter periods when the soil becomes sufficiently dry to develop desiccation cracks and concentration of turbation effects in the active layer above the permafrost table.

Table 4. Morphological description of some northern clayey soils.

Horizon	Depth (cm)	Colour (moist)	Texture*	Structure+	Special Features
	1. Solonetzic Gray Luvisol 55° 32'N, 98° 03'W				
L, F, H	6-0	5YR 2.5/2	—	—	
Ae	0-7	10YR 4/2	C	gr	
AB	7-10	10YR 5/3	C	sbk	vertical cracking
Btnj	10-17	10YR 4/2.5	HC	sbk	vertical cracking
Bt	17-30	10YR 4/2	HC	gr	
BC	30-36	10YR 5/2	HC	gr	
Ckgj	36-100	10YR 4.5/3	HC	mss	
	2. Solonetzic Gray Luvisol 55° 55'N, 97° 42'W				
L, H	2-0	10YR 2/1	—	—	
Ae	0-6	10YR 5/2	SiC	gr	
AB	6-16	10YR 4/3	C	pl	vertical cracking
Btnj	16-34	10YR 4/2	SiC	abk	vertical cracking
Bt	34-50	10YR 4/3	SiC	sbk	
Ck1	50-98	10YR 4/4	C	sbk	
Ck2	98-138	10YR 4/4	HC	gr	
	3. Orthic Turbic Cryosol 57° 50'N, 99° 36'				
L, F	6-0	10YR 5/4	—	—	
Bmy1	0-5	10YR 5/2	SiCL	gr	strongly cryoturbated
Bmy2	5-12	2.5Y 4/2	SiCL	sbk	strongly cryoturbated
Cgjy1	12-30	2.5Y 3.5/2	SiCL	mss	strongly cryoturbated
Cgjy2	30-60	2.5Y 3.5/2	SiCL	mss	strongly cryoturbated
Cgjy3	60-90	2.5Y 3.5/2	SiC	mss	strongly cryoturbated
Cz	90-98	2.5Y 3.5/2	SiCL	mss	
	4. Gleyed Gray Luvisol 48° 32'N, 81° 04'				
Ap	0-13	10YR 2/1	SiCL	sbk	
AB	13-18	10YR 5/2	SiCL	sbk	
Btgj	18-40	10YR 4/3	C	sbk	
Ckg	40+	10YR 6/3	SiCL	sbk	

*Texture: C-clay; HC-heavy clay; SiC-silty clay; SiCL-silty clay loam
+Structure: pl-platy; abk-angular blocky; sbk-subangular blocky; gr-granular; mss-massive

Present Use

The vast majority of northern clayey soils remain in their native state. Better drained clayey soils in the North Boreal Ecoclimatic Province support commercial stands of coniferous and mixed-wood forests. The extent of commercial forests, however, decreases rapidly with increasing latitude. Forest and other vegetation also have great value as wildlife habitat. Hunting and trapping are of both commercial and social value to many native people.

Small areas of clayey soils have been cleared and broken for agriculture. Cereal crops, forage, and potatoes have been grown with varying success due to climatic conditions. Potential for vegetable production has been proven by the operation of some market gardens developed by native communities and individuals for private use. Townsite development and associated infrastructure such as road construction have occurred on northern clayey soils, as well. If no permafrost is present, problems encountered are similar to those experienced in the south. Where permafrost exists, however, long-term problems with roadbed subsidence and shifting of building foundations can be expected if appropriate design measures are not taken.

Table 5. Some selected physical and chemical characteristics of some northern Manitoba and Ontario clayey soils.

Horizon	Depth (cm)	Total sand	Silt (%)	Clay (%)	pH	Cond. (mmhos/cm)	C	CEC (%)	Exchangeable Cations (meq) Ca	Mg	meq/100 g K	Na
	1. Solonetzic Gray Luvisol 55° 32'N, 98° 03'W											
L, F, H	6-0	—	—	—	5.1	47.6		103	40.0	14.4	1.6	0.1
Ae	0-7	5	39	56	5.5	5.7		46	22.5	10.7	0.8	0.1
AB	7-10	3	31	66	5.7	2.0		33	17.1	9.2	0.8	0.1
Btnj	10-17	1	13	86	6.1	1.4		34	18.3	10.3	0.9	0.2
Bt	17-30	1	6	93	6.5	0.7		34	20.7	10.5	0.9	0.2
BC	30-36	1	5	94	7.4	—	1.1	35	25.8	9.9	0.8	0.2
Ckgj	36-100	1	19	80	7.6	—	11.1	27	—	—	—	—
	2. Solonetzic Gray Luvisol 55° 55'N, 97° 42'W											
L, H	2-0	—	—	—	5.9	32.4		87	50.5	8.3	1.7	0.4
Ae	0-6	3	39	58	5.5	4.0		38	16.5	4.4	0.6	0.1
AB	6-16	1	38	61	5.6	2.5		31	15.8	4.3	0.6	0.1
Btnj	16-34	1	46	53	5.9	1.0		35	18.9	7.2	1.0	0.2
Bt	34-50	0	54	46	7.0	0.5		34	19.5	6.5	1.1	0.2
Ck1	50-98	0	40	60	7.6	—	8.3	31	18.8	6.1	0.8	0.2
Ck2	98-138	0	31	69	7.7	—	7.7	31	30.8	5.9	1.0	0.2
	3. Orthic Turbic Cryosol 57° 50'N, 99° 36'W											
L, F	6-0	—	—	—	3.5	36.6		67	10.2	6.2	1.8	0.6
Bmy1	0-5	6	64	30	4.2	6.2		29	3.7	4.8	0.5	0.3
Bmy2	5-12	7	61	32	4.8	3.8		20	8.6	5.0	0.5	0.2
Cgjy1	12-30	8	61	31	5.7	1.6		15	10.5	5.4	0.4	0.2
Cgjy2	30-60	7	59	34	5.8	—		16	12.4	5.5	0.4	0.2
Cgjy3	60-90	6	53	41	6.5	—		17	13.5	5.6	0.5	0.2
Cz	90-98	7	5	38	6.6	—		19	13.5	5.5	0.5	0.2
	4. Gleyed Gray Luvisol 48° 32'N, 81° 04'											
Ap	0-13	8	44	48	7.0	3.0		—	—	—	—	—
AB	13-18	7	50	43	7.1	0.9		—	—	—	—	—
Btgj	18-40	2	27	72	7.7	0.7		—	—	—	—	—
Ckg	40+	1	44	55	7.9	—		—	—	—	—	—

Some Properties of Clayey Soils in British Columbia and Northern Alberta

Almost all of British Columbia lies within the Cordilleran Ecoclimatic Provinces, which include the Pacific Cordilleran (referred to as the "coast"), the South Interior Cordilleran (referred to as the "southern interior"), and the North Interior Cordilleran (referred to as the "northern interior") Ecoclimatic Provinces (Figure 1). The extreme northeastern corner of British Columbia, together with northern Alberta, lies within the North Boreal Ecoclimatic Province. For more information on the North Boreal Ecoclimatic Province, see the preceeding section.

Morphology

The upper solum of well drained coastal clayey soils (Soil 1, Table 6) generally has subangular blocky structure, while that of the poorly drained coastal clayey soils has a granular structure. The well drained clayey soils in the southern interior have granular to subangular blocky surface soil structures. The clayey Luvisolic soils in the northern interior (Soil 6, Table 6), however, have platy surface soil structures. The B horizons of all of these clayey soils have prismatic structures which break to subangular or angular blocky. The structures of all of these B horizons are very pronounced and well developed.

Cracks approximately 1 to 2 cm wide and extending to a depth of 1 m or more are common in the clayey soils of the southern interior (Soils 3, 4 and 5, Table 6). During very dry summers, the coastal clayey soils (Soils 1 and 2, Table 6) also develop similar cracks. These cracks, however, are less visible because of the friable surface materials. Cracking occurs less frequently in the northern interior clayey soils (Soil 6, Table 6) than in the soils of the southern interior. This lack of cracking is probably due to the organic surface horizons (forest litter) and the usually higher moisture levels. Slickensides have been noted in the interior clayey soils, but none have been reported on the coast.

Physical Properties

The texture of these soils is silty clay to clay, with the highest clay content usually occurring in the B horizon (Table 7). The upper horizons of these soils are usually coarser textured because of eluviation. The low permeability results in periodically perched water tables, especially during the winter months in the coastal clayey soils. Perched water tables also occur in the northern interior clayey soils because of seasonal frost in the subsoil. This frost remains well into June, especially in sites having a dense moss cover.

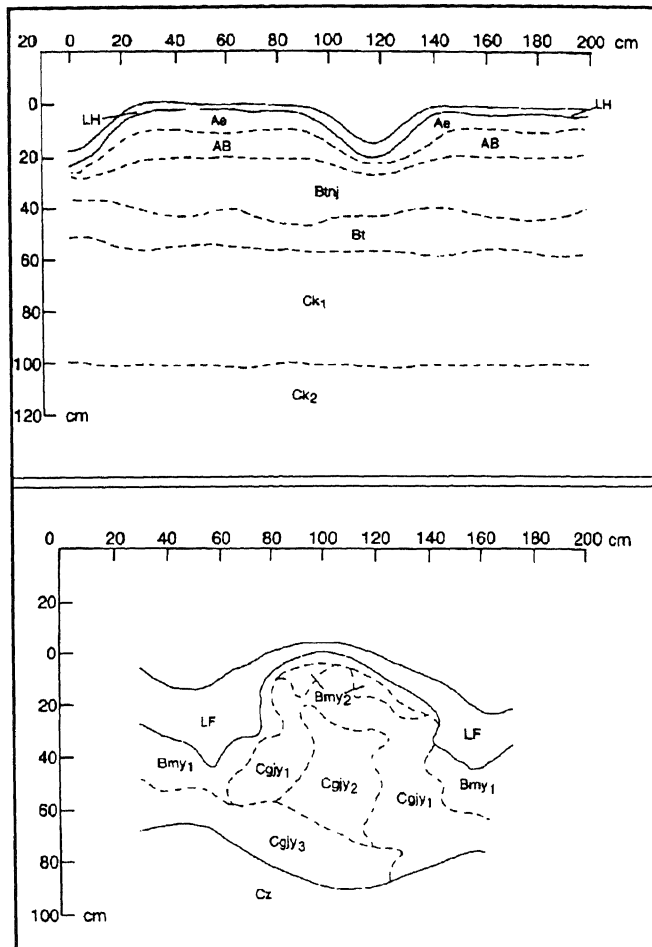


Figure 13. a. Cross section through subdued microhummocks and continuous soil horizons. Soil 2, Table 4. Solonetzic Gray Luvisol, 63P22. Location: Manitoba 55° 55'N, 97°o 42'W .b. Cross section of an earth hummock with strongly mounded microrelief and cryoturbated horizons. Soil 3, Table 4. Orthic Turbic Cryosol, 64G1. Location: Manitoba 57° 50'N, 99° 36'W.

The clay mineralogy of the coastal clayey soils is dominated by vermiculite and chlorite, in both the coarse and fine clay fractions. Smaller quantities of smectite and mica also occur. The interior clayey soils are dominated by smectite and vermiculite.

Chemical Properties

Both the coastal and northern interior clayey soils have strongly to very strongly acid surface mineral horizons and slightly acid to mildly alkaline parent materials (Table 7). The southern interior clayey soils, on the other hand, are slightly acid to neutral, with mildly to moderately alkaline parent materials.

The organic carbon content of the surface mineral horizons of well to imperfectly drained clayey soils is about 2%, with coastal clayey soils having slightly higher contents than those in the interior. The poorly drained clayey soils (Soil 2) generally have higher carbon contents than do the imperfectly and well drained soils (Table 7).

The cation exchange capacity varies from 11 to 45 meq/100g. These values are largely dependent on the soil horizon, clay mineralogy, and organic matter content.

Table 6. Morphological description of some British Columbia clayey soils.

Horizon	Depth (cm)	Colour (moist)	Texture*	Structure+	Special Features
	1. Orthic Dystric Brunisol 49° 17'N, 124° 54'W				
L	8-3	—	—		
F, H	3-0	—	—	—	
Bmcc	0-6	5YR 5/6	L	gr	iron-manganese concretions
Bm1	6-18	7.5YR 4/3	L	gr	
Bm2	18-25	10YR 4/3	SiC	gr	
Bm3	25-31	2.5Y 4/4	SiC	gr	
Bm4	31-50	2.5Y 4/4	SiL	sbk	
BC	50-75	2.5Y 5/2	L	pl	
C1	75-96	2.5Y 5/2	L	mss	
C2	96-127	2.5Y 4/4	L	mss	
IIC	127+	2.5Y 4/2	SL	sg	
	2. Humic Luvic Gleysol 49° 06'N, 122° 37'W				
Ap	0-15	10YR 2/2	SiC	gr	
Aeg	15-25	2.5Y 5/2	SiL	sbk	
AB	25-32	2.5Y 4/2	SiCL	pr	
Btg1	32-48	2.5Y 4/2	SiC	pr	
Btg2	48-70	5Y 4/3	SiC	pr	
BC	70-92	5Y 4/1	SiC	pr	
Cg1	92-120	5Y 4/1	-	mss	
Cg2	120+	5Y 4/1	-	mss	
	3. Orthic Dark Grey Chernozem 50° 25'N, 119° 11'W				
Ap1	0-21	10YR 4/2	C	gr	
AB	21-30	10YR 5/2	HC	gr	
Bt	30-60	10YR 4/3	HC	pr	
BCca1	60-93	10YR 5/4	HC	abk	
BCca2	93-115	2.5Y 4/2	HC	abk	
Ck	115-135	2.5Y 4/2	HC	mss	
	4. Eluviated Dark Brown Chernozem 49° 49'N, 119° 37'W				
Ap	0-14	10YR 5/2	SiC	abk	
Ae	14-20	10YR 5/2	SiC	pl	
Bt	20-35	10YR 5/2	SiL	pr	
BC	35-49	2.5Y 5/2	SiC	pr	
Ck	49-82	5Y 7/2	HC	pl	
	5. Orthic Gray Luvisol 50° 26'N, 119° 11'W				
Ap	0-9	10YR 3/2	SiL	sbk	
Ae	9-24	10YR 5/2	SiCL	abk	
Bt1	24-39	10YR 4/2	C	pr	
Bt2	39-64	10YR 5/2	C	pr	
BCca	64-90	10YR 6/2	CL	abk	
Ck	108-125	10YR 4/2	CL	pl	
	6. Orthic Gray Luvisol 54° 23'N, 124° 16'W				
L, F, H	5-0	10YR 5/2	—	—	
Ae	0-5	10YR 5/2	C	pl	
AB	5-13	10YR 3/3	HC	abk	
Bt1	13-25	10YR 3/3	HC	col	
Bt2	25-51	10YR 3/3	HC	pr	
BC	51-62	10YR 3/3	HC	pr	
C	62-110	10YR 3/3	HC	—	

*Texture: C-clay; HC-heavy clay; SiC-silty clay; SiCL-silty clay loam; SL-sandy loam; CL-clay loam; L-loam
+Structure: pl-platy; abk-angular blocky; sbk-subangular blocky; gr-granular; mss-massive; pr-prismatic; col-columnar; sg-single grain

Present Use

The clayey soils in British Columbia and northern Alberta are used for agriculture and forestry. Along the coast, most of these soils are used for pasture and hay production and for forestry, although in the lower Fraser River Valley they also are used for growing cereal crops. In the southern interior almost all clayey soils are used for agriculture and are among the best agricultural soils for growing hay, cereal grains and, where the climate is suitable, tree fruits. In the northern interior, these soils are used for forage production, pasture, and forestry.

The main limiting factors for use are the shallow rooting depth, the slow warming in spring, the maintenance of the surface organic matter content in the northern interior, and the poor trafficability because of the periodically perched water tables. There are also problems in maintaining the soil structure and limiting compaction.

Classification of Turbated Soils

Both cryoturbation and shrink-swell processes in soil result in physical disturbance affecting the entire soil system. Soils characterized by such turbation processes occur under environments extending from the arctic to the tropics.

In the Canadian soil classification, soils associated with permafrost are classified as Cryosols (Agriculture Canada Expert Committee on Soil Survey, 1987) while in the American soil taxonomy such soils are recognized only at the subgroup level in the various soil orders. All textural classes of these permafrost soils commonly are affected by cryoturbation processes resulting in churning of soil material and development of various kinds of patterned ground. Additional cryogenic forces are responsible for the formation of cracking in these soils. The morphology and properties of clayey soils associated with permafrost are provided in the foregoing description of the Arctic, Subarctic, and North Boreal Ecoclimatic Provinces.

In the American soil taxonomy, strongly turbated soils south of the permafrost region are

classified as Vertisols (Soil Survey Staff, 1975). These soils have developed in clayey materials with high shrink-swell potential, resulting in both the movement of soil materials and cracking. The volume change required for vertic soil properties to develop is greatest under environments which permit greatest oscillation between wet and dry soil moisture content (Wilding and Tessier, 1988).

Soils affected by shrink-swell processes are recognized in *Soil Taxonomy* (Soil Survey Staff, 1975) at the order level as Vertisols. In contrast, similar degrees of physical disturbance resulting from cryogenic forces in cold soils are recognized at the subgroup level in various soil orders. Similarity in degree of physical disturbance and soil cracking suggests that the logic of this ordering of soil properties should be reconsidered in any future revision of *Soil Taxonomy*. Establishment of an order of soils (equivalent to the Canadian Cryosolic Order) to recognize the morphology resulting from cryogenic forces associated with permafrost would correspond to recognition of shrink-swell soils occurring in southern regions in the Vertisol Order. Dr. Guy Smith (1986, p. 61) suggested that establishment of such an order should be considered by a small international committee.

Table 7. Physical and chemical characteristics of some British Columbia clayey soils.

Horizon	Depth (cm)	Total sand	Silt (%)	Clay (%)	pH	Cond. (mmhos/cm)	C	CEC (%)	Exchangeable Cations (meq) Ca	Mg	meq/100 g K	Na
	1. Orthic Dystric Brunisol 49° 17'N, 124° 54'W											
L	8-3	—	—	—	6.0	—	39.4	96	78.1	7.7	2.0	0.3
F, H	3-0	—	—	—	5.0	—	40.3	101	59.0	16.3	1.4	0.3
Bmcc	0-6	31.1	41.3	27.6	4.7	—	3.3	27	1.8	0.4	0.1	0.1
Bm1	6-18	24.4	47.8	27.7	4.7	—	2.4	26	1.7	0.4	0.1	0.1
Bm2	18-25	15.7	49.3	35.0	4.8	—	2.3	27	4.6	0.6	0.1	0.1
Bm3	25-31	10.5	51.9	37.6	4.9	—	1.5	25	7.8	0.9	0.1	0.1
Bm4	31-50	11.7	48.2	40.0	5.2	—	0.9	33	17.6	1.8	0.1	0.1
BC	50-75	40.5	38.8	20.7	5.8	—	0.3	34	22.2	1.7	0.1	0.1
C1	75-96	46.6	35.6	17.8	6.4	—	0.3	32	22.4	1.5	0.1	0.1
C2	96-127	37.0	44.4	18.6	6.1	—	0.3	29	19.2	1.2	0.1	0.1
IIC	127+	65.5	32.4	2.0	6.9	—	0.1	4	2.0	0.1	0.0	0.0
	2. Humic Luvic Gleysol 49° 06'N, 122° 37'W											
Ap	0-15	2.0	64.0	34.0	5.6	—	6.7	36	5.6	2.6	0.3	0.5
Aeg	15-25	2.0	70.0	28.0	5.6	—	0.7	18	4.3	4.6	0.1	0.3
AB	25-32	5.0	56.0	38.0	5.9	—	0.5	26	7.3	10.1	0.1	0.6
Btg1	32-48	1.0	50.0	49.0	6.7	—	0.4	35	10.5	18.4	0.2	2.2
Btg2	48-70	1.0	49.0	50.0	7.8	—	0.2	35	10.1	16.4	0.2	4.3
BC	70-92	1.0	49.0	50.0	7.9	—	—	32	9.9	13.7	0.3	4.1
Cg1	92-120	1.0	49.0	50.0	7.8	—	—	31	—	—	—	—
Cg2	120+	—	—	—	8.0	—	—	25	8.1	14.3	0.4	4.1
	3. Orthic Dark Gray Chernozem 50° 25'N, 119° 11'W											
Ap	0-21	8.2	—	58.0	6.0	—	3.0	43	7.7	11.5	1.8	—
AB	21-30	5.0	—	72.2	6.4	—	2.2	50	16.1	16.4	1.9	0.3
Bt	30-60	5.0	—	74.5	6.7	—	2.0	—	—	—	—	—
BCca1	60-93	0.0		90.0	7.1	—	0.7	—	—	—	—	—
BCca2	93-115	1.0		88.5	7.3	—	0.5	—	—	—	—	—
Ck	115-135	1.0		94.2	7.5	—	0.2	—	—	—	—	—
	4. Eluviated Dark Brown Chernozem 49° 49'N, 119° 37'W											
Ap	0-14	16.8	47.7	35.5	6.8	—	3.8	32	19.6	6.2	2.9	0.1
Ae	14-20	19.1	49.3	31.5	6.7	—	1.6	26	14.9	6.3	1.8	0.1
Bt	20-35	7.6	43.0	49.6	6.7	—	0.8	33	20.3	10.0	1.3	0.1
BC	35-49	1.4	66.2	32.4	7.0	—	0.7	29	18.5	11.9	0.8	0.1
Ck	49-82	1.6	38.4	60.0	7.7	—	—	31	18.9	11.6	0.6	0.3
	5. Orthic Gray Luvisol 50° 26'N, 119° 11'W											
Ap	0-9	10.2	31.7	58.0	7.4	0.13	4.8	—	—	—	—	—
Ae	9-24	5.2	25.7	69.0	7.4	0.13	2.7	—	—	—	—	—
Bt1	24-39	2.0	17.0	81.0	7.6	0.08	1.3	—	—	—	—	—
Bt2	39-64	0.0	14.5	85.5	8.2	0.23	0.9	—	—	—	—	—
BCca	64-90	0.0	16.0	84.0	8.2	2.60	0.5	—	—	—	—	—
CBca	90-108	0.0	10.0	90.0	8.1	4.70	0.5	—	—	—	—	—
CK	108-125	3.0	55.5	41.5	8.1	4.50	0.5	—	—	—	—	—
	6. Orthic Gray Luvisol 54° 23'N, 124° 6'W											
L, F, H	5-0	—	—	—	5.4	—	24.6	104	39.6	27.8	4.3	0.3
Ae	0-5	—	—	—	4.9	—	2.6	36	8.3	11.1	0.8	0.1
AB	5-13	7.0	27.0	66.0	4.8	—	1.4	27	5.7	10.3	0.5	0.1
Bt1	13-25	1.0	16.0	83.0	4.5	—	1.0	37	8.4	16.7	1.1	0.2
Bt2	25-51	1.0	14.0	85.0	4.5	—	1.1	43	9.4	23.0	0.7	0.7
BC	51-62	—	—	—	4.7	—	—	37	8.2	27.2	0.5	0.9
C	62-110	—	31.0	69.0	5.6	—	—	31	7.5	21.4	0.3	1.3

Summary

Representative clayey soils from the Arctic, Subarctic, North Boreal, and Cordilleran Ecoclimatic Provinces were evaluated in order to determine their properties and classification.

All clayey soils in the Arctic and Subarctic Ecoclimatic Provinces are associated with permafrost and earth hummocks and are classified as Cryosols. Their morphology is dominated by cryoturbated features and the occurrence of perennially frozen soil horizons. If these soils occur on sloping topography, they are subject to solifluction and erosion when the thermal balance is disturbed. Cracking is common in these soils, but these cracks are the result of thermal contraction, not desiccation. Clayey soils occurring on level topography are subject to thermokarst.

Well to imperfectly drained clayey soils occurring in the North Boreal Ecoclimatic Province have either a brownish Bm horizon without significant clay accumulation or a Bt horizon with clay accumulation. Those soils having Bm horizons are classified as Brunisols, those with Bt horizons, as Luvisols. Poorly drained clayey soils in this region commonly have a peaty surface horizon and strongly gleyed and mottled

mineral horizons. These soils are classified as Gleysols or, if permafrost is present, as Cryosols.

In the South Interior Cordilleran Ecoclimatic Province, those clayey soils having a dark L-F-H or organic-rich Ah (or Ap) horizon are classified as Luvisols or Chernozems. When characterized by pronounced mottles or matrix colours of low chroma, they are classified as Gleysols.

Arctic and subarctic clayey soils show no evidence of slickensides and desiccation cracking. These soils, with the exception of the surface horizons, are moist to wet throughout the summer because the cool climate results in lower evapotranspiration than is found in the warmer southern soils. The melting of ice in both the seasonally frozen layer and the near-surface permafrost continuously releases moisture during the summer months, keeping these soils moist or wet. Slickensides have been observed in some of the clayey Luvisols in the boreal (Manitoba and Saskatchewan) and Cordillera (southern interior of British Columbia). These soils also show evidence of cracking due to desiccation, although no grumic or churning characteristics have been observed.

According to the present criteria, some of the clayey soils may fit some of the Vertisol suborders of the American soil taxonomy. At the present time, however, there is not sufficient data to verify this. Vertic properties, if present, do not appear to be strong enough to warrant the soils being classified under the proposed Cryert suborder.

References

Acton, D., R. Smith and G. Coen. These proceedings. Clayey soils of the grasslands and southern boreal.

Agriculture Canada Expert Committee on Soil Survey. 1987. The Canadian system of soil classification. Research Branch, Agriculture Canada, Publication No. 164, second edition, 164 p.

Atmospheric Environment Service. 1982a. Canadian climate normals, temperature, and precipitation: the North - Yukon Territory and Northwest Territories. Environment Canada, Atmospheric Environment Service, UDC:551.582(712), 55 p.

Atmospheric Environment Service. 1982b. Canadian climate normals, temperature, and precipitation: Prairie Provinces. Environment Canada, Atmospheric Environment Service, UDC:551.582(712), 429 p.

Bentley, C.F. (editor). 1978. Photographs and descriptions of some Canadian soils. Volume 4, Eleventh Congress of the International Society of Soil Science, Edmonton, Canada, 98 p.

Bostock, H.S. 1970. Physiographic subdivisions of Canada. in Geology and Economic Minerals of Canada. Department of Energy, Mines and Resources Canada, Geological Survey of Canada, Economic Geology Report No. 1, p. 9-30.

Brown, R.J.E. 1967. Permafrost in Canada. Geological Survey of Canada, Department of Energy, Mines and Resources, Map 1246A, First Ed.

Brown, R.J.E. and W.O. Kupsch. 1974. Permafrost terminology. National Research Council of Canada, NRCC 14274, Technical Memorandum No. 111, 62 p.

Clayton, J.S., W.A. Ehrlich, D.B. Cann, J.H. Day and I.B. Marshall. 1977. Soils of Canada. Research Branch, Canada Department of Agriculture, 2 vols. and accompanying maps.

Ecoregion Working Group. 1989. Ecoclimatic regions of Canada, first approximation. Ecoregion Working Group of the Canada Committee on Ecological Land Classification. Ecological Land Classification Series 23, Sustainable Development Branch, Canadian Wildlife Service, Conservation and Protection, Environment Canada, Ottawa, Canada, 119 p. and map.

Harris, S.A., H.M. French, J.A. Heginbottom, G.H. Johnston, B. Ladanyi, D.C. Sego and R.O. van Everdingen. 1988. National Research Council of Canada Technical Memorandum No. 142, NRC Report No. 27952, 156 p.

Mackay, J.R. 1972. Some observations on ice-wedges, Garry Island, NWT. in Kerfoot, D.E., ed. Mackenzie Delta area monograph, 22nd International Geological Congress, Montreal, Brock University, St. Catharines, Ontario, 174 p.

Mackay, J.R. 1974. Ice-wedge cracks, Garry Island, Northwest Territories. Canadian Journal of Earth Science, 11:13361383.

Mills, G.F., R.G. Eilers and H. Veldhuis. These proceedings. Thermal regime and morphology of clayey soils in Manitoba, Canada.

Pettapiece, W.W., C. Tarnocai, S.C. Zoltai and E.T. Oswald. 1978. Guidebook for a tour of soil, permafrost and vegetation relationships in the Yukon and Northwest Territories of northwestern Canada, Tour 18. 11th Congress of the International Society of Soil Science, Edmonton, Canada, 165 p.

Prest, V.K. 1970. Quaternary geology. in Geology and Economic Minerals of Canada. Department of Energy, Mines and Resources Canada, Geological Survey of Canada, Economic Geology Report No. 1, p. 675-764.

Smith, G.D. 1986. The Guy Smith interviews: Rationale for concepts in soil taxonomy. T.R. Forbes (editor). SMSS Technical Monograph No. 11, Washington, D.C., 259 p.

Soil Survey Staff. 1975. Soil taxonomy. Soil Conservation Service, U.S. Department of Agriculture, Agriculture Handbook No. 436, Washington, D.C., 754 p.

Tarnocai, C. 1978. Distribution of soils in northern Canada and parameters affecting their utilization. 11th International Congress of Soil Science Transactions, Volume 3, Symposia Papers, p. 281-304.

Tarnocai, C. and S.C. Zoltai. 1978. Earth hummocks of the Canadian arctic and subarctic. Arctic and Alpine Research 10:581-594.

Washburn, A.L. 1979. Geocryology. Edward Arnold, London. 406 p.

Wilding, L.P. and D. Tessier. 1988. Genesis of Vertisols: shrink-swell phenomena. In: Vertisols: Their distribution, properties, classification and management. L.P. Wilding and R. Puentes (editors), Texas A&M University System and Soil Management Support Services, College Station, Texas, p. 5581.

Discussion

Don Gross - What part of the soil has the thermal properties to expand and contract to form the thermal contraction cracks? Does the mineral fraction have such thermal contraction properties?

C. Tarnocai - Frost cracking is the result of thermal contraction of the soil in subfreezing temperatures. When the frost crack develops, the soil is in a frozen state. That is, the active layer (the layer which thaws and refreezes annually) is frozen to the permafrost. The initial frost crack starts at the surface of the soil and can extend to a depth of 4 m. The cracks are primarily imprinted in the permafrost rather than in the active layer and, thus, they recur at the same place year after year. Thermal contractions (frost cracks) develop in both mineral and organic soils.

J.L. Sehgal - The arctic and subarctic soils of northern Canada have been grouped as CRYOSOLS. We have such soils in the high Himalayas which qualify as Cryochrepts. How would you classify these soils in the U.S. System of *Soil Taxonomy*?

C. Tarnocai - I am not experienced in classifying permafrost soils using the U.S. *Soil Taxonomy*. We tried to use this soil classification in the early 1970s, but had great difficulty. The northern soils classified as Cryosols in the Canadian System of Soil Classification fit into the Pergelic subgroups (e.g. Pergelic Cryochrept and Pergelic Ruptic Cryochrept) of the U.S. *Soil Taxonomy*.

Warren Lynn - You mentioned that frozen soils cracked at weak points. What are the weak points and can they be predicted?

C. Tarnocai - The weak point could be the change in ice content or textural differences. There is no definite way, as far as I know, to predict these weak points. If a frost crack develops, however, the soil will crack at the same place in the future.

Jim Richardson - Please cover the mechanism for frost cracking?

C. Tarnocai - Frost cracking is the fracturing of soil material by thermal contraction in subfreezing temperatures. All mineral soil materials contract a certain amount at low temperatures but; since ice is the critical material in frozen soils, its contraction plays an important role. Pure ice has a coefficient of linear contraction of about $45x10^{-6}$ °C-1 at -40°C. Soil material, however, is non-homogeneous and thus a complex system. Other factors affecting cracking are the salt content, the highly variable distribution of ice in the frozen soil, and the insulating effect of the snow. Frost cracks can develop in both coarse and fine textured soil materials as well as in organic materials. Fine textured, moisture-rich soils are probably the soils which are most susceptible to frost cracking.

A. Osman - What is the origin of the sodium salts in the soils of northern Canada and how are they formed?

C. Tarnocai - The salts in the northern soils originate from the parent materials. Soils which developed in residual or till materials derived from shale and marine deposits are commonly associated with salts. A whitish salt crust can develop on the surface of these soils as a result of evaporation of moisture from the soil surface during warmer summer days.

Vertisols of France

Daniel Tessier[1], Ary Bruand[2], and Yves-Mary Cabidoche[3]

Abstract

This paper deals with a general presentation of Vertisols and related soils of France. Vertisols and related soils are mainly developed on clayey geological substrata. Clay minerals are either 2:1 interstratified illite-smectite clays or smectite-rich materials. Shrinkage curves on undisturbed samples exhibit a normal shrinkage in a wide range of water content. The shrinkage amplitude is highly variable and appeared to be dependent on both clay content and mineralogy. Water retention curves differed considerable from one soil to another. A statistical study showed the respective contribution of clay content, clay mineralogy (CEC and EGME surface areas), and soil fabric to soil physical properties.

Introduction

The geographic situation and climatic conditions encountered in France are not generally favorable to the development of Vertisols. However, some soils found on a large range of clay-rich parent materials can be classified following *Soil Taxonomy* in the order of Vertisols (Duchaufour, 1982). Moreover, the behavior of many other related soils is strongly influenced by the presence of vertic B horizons, although these soils are not classified as Vertisols. This is particularly the case for Pelosols such as defined in the German classification (Bonneau et al., 1965 and 1967; Begon and Jamagne, 1973). Because these vertic B horizons play an essential role in soil behavior, numerous research works have been aimed at establishing a relationship between intrinsic characteristics such as clay mineralogy, texture, and behaviors.

This paper first presents the geographic situation, the climatic conditions, and the distribution of Vertisols and related soils in relation to the nature of parent materials. It then examines the main properties of these soils, especially their physical properties.

Geography and Climate

France extends from 42° to 52° north latitude and from 5° and 8° west and east longitude, respectively. The characteristics of the climate prevailing over Vertisols and related soils in France are presented in Fig. 1. Except for mountain areas, rainfall ranges from 600 to 900 mm per year. In the north-west part of the country, rainfall is distributed regularly. The oceanic temperate climate is characterized by mild temperatures, and 7°C and 18°C are the mean temperatures in January and July, respectively, with 180 days of rainfall in the Brest area. Further eastward the climate changes from oceanic to continental. Winters are dry and cold, whereas summers are wet and warmer. In the southern part of France the rainfall distributed irregularly. The dry season is well expressed, with about 60 rainfall days per year, while mean temperatures reach 7°C and 23°C for the colder and warmer months, respectively, i.e., January and July. The general elevation of Vertisols and related soils ranges from sea level to 500 m above sea level.

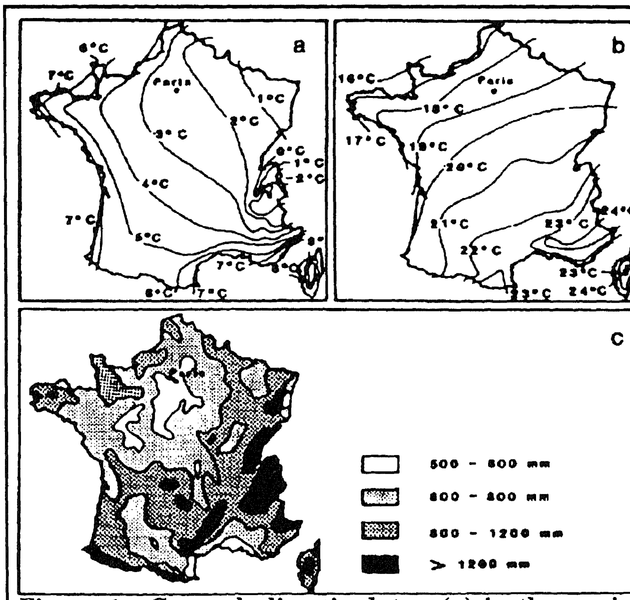


Figure 1. General climatic data: (a) isotherms in January; (b) isotherms in July; and (c) mean annual precipitation.

[1]Station de Science du Sol, INRA, Route de S[t] Cyr, 78026 Versailles, France.

[2]SESCPF, INRA, Centre d'Orleans, Ardon, 45160 Olivet, France.

[3]CRAAG, INRA, Centre De Guadeloupe, Petit Bourg, BP 1232 97184 Pointe a Pitre, Guadeloupe, Frenc Indies.

Spatial Distribution

Vertisols and related soils (i.e., those with vertic B horizons) occur in France on areas with intensive agricultural activities. They cover about 5 percent of cultivated soils and are located mainly in the Paris and Aquitanian Basin (Fig. 2). Vertisols were described in the southern part of the Beauce Plain near Orleans by Fedoroff and Fies (1968), Horemans (1984), and Arrouays (1987). In the south of France, Bonfils (1988) found Vertisols close to the Mediterranean Sea. Vertisols in Guadeloupe the French West Indies have developed on the coastal plain under climatic conditions very different from those in metropolitan France (Jaillard and Cabidoche, 1984).

Parent Materials

Vertisols occurring on basalts are present near Agde in the south of France under Mediterranean climate (Bonfils, 1988). Other Vertisols have developed on clayey sedimentary deposits which are usually interstratified illite-smectite or smectite-rich materials of both secondary and tertiary ages (Bornand et al., 1975 and 1984; Jamagne et al., 1970; Fedoroff and Fies, 1968). The more frequent situation corresponds to the presence of clay minerals similar to those present at the time of the sediment formation.

Clay minerals are sometimes transformed during pedogensis, as described by Robert et al., (1973). For instance, in the Paris Basin the Albian and Cenomanian glauconite-rich materials are weathered into iron-rich smectites, and Vertisols have developed in these smectites (Isambert, 1984). Soils related to Vertisols appear on numerous sedimentary clayey outcrops. These soils are characterized by vertic structured B horizons. Parent materials are sometimes clayey marls on Trias and Lias sediments (Nguyen Kha, 1973; Bonneau et al., 1967), on glauconitic Cenomanian formation (Isambert, 1984), or on clayey Quaternary sediments (Salin, 1983). According to the type of sedimentation and the importance of diagenesis, parent clay minerals are either 2:1 interstratified illite-smectite clays or smectitic-rich materials.

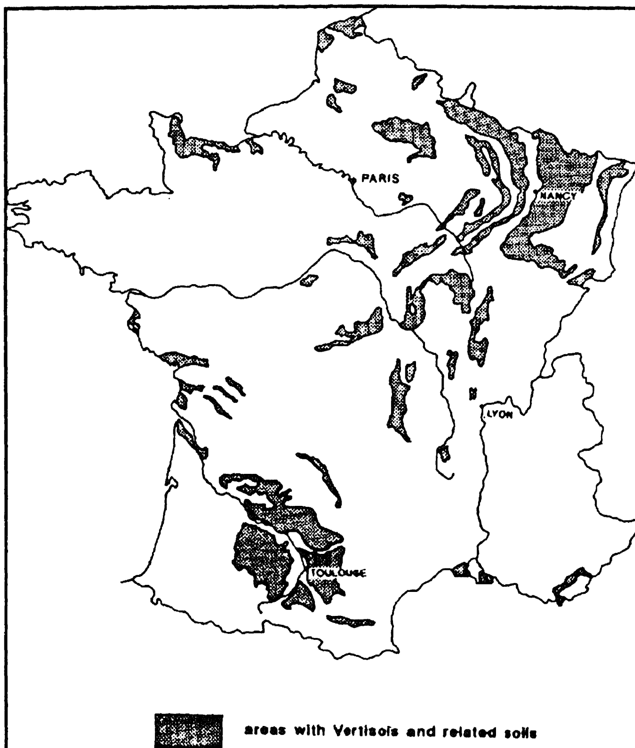


Figure 2. Schematic map of Vertisols and related soils with both A and B clayey horizons.

Particle Size, Mineralogy, and Chemistry

Table 1 summarizes the main characteristics of 9 B horizons belonging to Vertisols or related soils. The < 2µm fraction is higher than 50 percent. These horizons have developed upon noncalcareous materials or are the final stage of a complete decarbonation process. Carbon content is always low (less than 1 percent) and cation exchange capacities are saturated with Ca, Mg, K, and Na cations, with Ca being dominant.

Fig. 3 shows the main types of mineralogy encountered. Clays have been classified according to cation exchange capacities and accessible surface to ethylene glycol mononethyl (EGME). Both characteristics are generally very well correlated. X-ray diagrams and transmission electron micropscopy show that the proportion of

TABLE 1: Main characteristics of 9 vertic B horizons and related soils

Soil	Organic Depth cm	Clay < 2µm	Silt 2-50µm	Sand 50µm-2mm	pH in water	$CaCO_3$ %	C E C $m.e.g^{-1}$	Carbon %
ST FRANCOIS	60-90	85.4	9.4	5.2	8.0	0.9	0.47	0.23
FREVILLE	24-44	82.6	14.2	3.2	8.2	0.5	0.43	0.67
BETHONVILLIERS	25-60	52.8	37.3	9.9	7.1	0.6	0.39	0.42
VILLERS-STONC.	25-45	66.2	30.6	3.2	7.5	0	0.35	0.55
THIFON-ALO	85-105	72.6	17.7	9.7	4.8	0	0.33	0.23
CHEZAL-BENOIT	60-95	92.9	4.5	2.6	4.7	0	0.26	0.19
LA BOUZOULE	20-50	56.0	42.2	1.8	8.1	0.9	0.20	0.71
LA VIAUDE	25-40	58.2	41.4	0.4	8.6	9.2	0.18	0.69
THIANGES	25-70	63.5	26.5	10.0	7.4	0	0.17	0.16

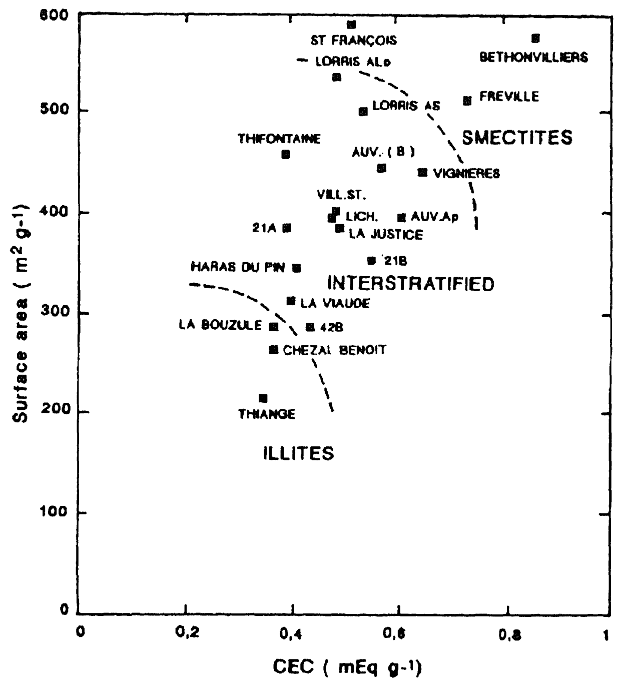


Figure 3. Cation exchange capacities and EGME surface areas as a function of clay mineralogy.

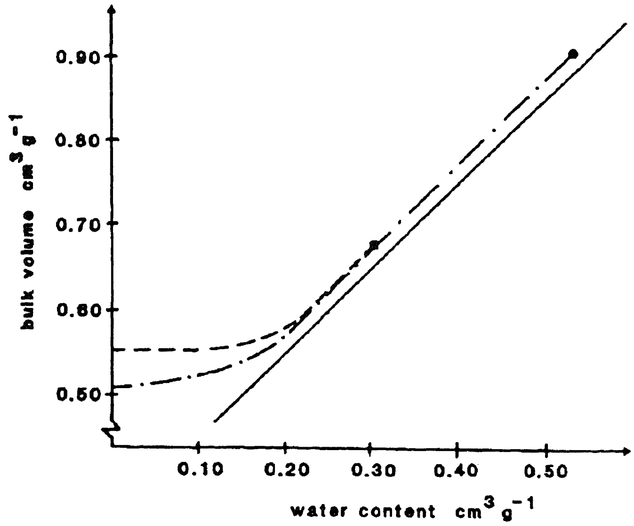


Figure 4. Shrinkage curves of two vertic B horizons with 1.1 g.cm-3 (-.-.-) and 1.5 g. cm-3 (—) bulk densities at field capacity, repectively.

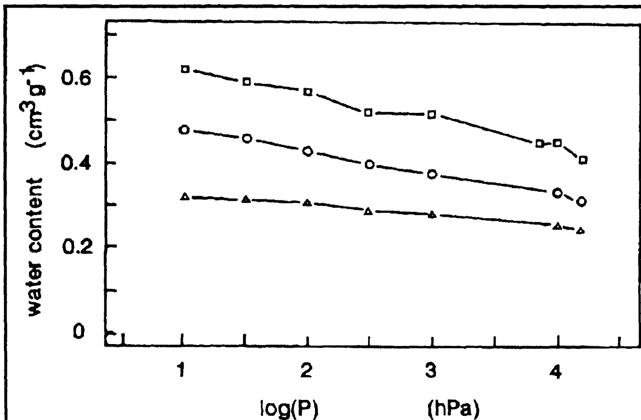


Figure 5. Water retention curves of three vertisols.

opened 2:1 layers ranges from about 30% to nearly 75%. Tessier et al. (1989) established that the higher the CEC the lower the clay particle size. consequences of shrink-swell phenomena, air entry point and structural stability were deduced. Thus, EGME and CEC appeared to be the best criteria to characterize soil clay mineralogy.

It was also noticed that soils developed on the same clay-rich parental material is mineralogically homogeneous and thus typical of the parent material.

Water Retention and Shrink-Swell Phenomena

These soil characteristics are generally considered in laboratory studies on centimetric clods (5 to 10 cm^3). To study them, large undisturbed samples are collected, preferably at the end of winter or at the beginning of spring (i.e., full and homogeneous wetting). Drying is avoided before measurements. Clods are obtained by hand breaking and equilibrated with a range of water potentials from -1 KPa to -2 MPa. Water contents and bulk densities are measured (Tessier and Berrier, 1979; Bruand and Prost, 1987). Similar measurements also can be made on clods in equilibrium with different relative humidities.

On the studied B horizons, drying curves exhibit a normal shrinkage in a wide water content range (Fig. 4). Normal shrinkage mainly corresponds to a complete water saturation. Nevertheless, Cabidoche and Voltz (1987) showed the presence of an incomplete water saturation due to biopores. On the other hand, water retention curves can be different to a large extent (Fig. 5). For instance, at a given water potential, water content can be twice as much in one vertisol as in another.

A device also was used to measure soil vertical movement (Cabidoche and Voltz, 1988). The results were interpreted in terms of horizontal and vertical cracking and water dynamics in biopores and clay matrix pores.

Relationship Between Measurable Intrinsic Properties and Behaviors

Vertisols and related soils of France have quite varied clay contents, clay mineralogies, water retention, and shrinkage curves. Attempts have been made to establish a relationship between measurable intrinsic properties and behaviors.

A study was carried out on the < 2µm clay fraction extracted from 8 vertisols and related soils (Tessier et al., 1990). Sample preparation was standardized: drying at 48% RH, slow rewetting up to a given water potential. A correlation between clay hydration properties and macroscopical clay swelling on one hand, EGME surface area and CEC on the other hand, was found.

These results were compared to water behaviors of clods sampled at field capacity (Bruand et al., 1988). The statistical relationship has shown the respective contribution of clay content, mineralogy, and fabric to soil properties in a range of water potentials from -32 KPa to -1.6 MPa (Bruand et al., 1988). Thus, it was established that taking into account soil CEC allows better prediction of water content than clay content does. At -32 KPa, the latter accounted for 58% of water content variance, in contrast to 79% for the former (Table 2). Similar results were obtained at -1.6 MPa.

Furthermore, the natural soil fabric also can be taken into account by measuring the pore volume due to the clay phase in the samples near field capacity. A simple way to take this parameter into account was to consider the bulk volume of the natural clods sampled at filed capacity. Pore volume is correlated positively with water content at different water potentials. This correlation allows 93 and 87% of the water content variance to be explained at -32 KPa and -1.6 MPa, respectively (Bruand et al., 1988). Similar results were obtained with shrinkage. The shrinkage amplitude appeared to be dependent on both clay content and mineralogy. The higher the bulk volume (or the lower the bulk density), the higher the shrinkage.

It was concluded that the prediction of soil physical properties is considerably improved by taking CEC into account. A further significant improvement was obtained when soil fabric, numerically expressed by the bulk volume, at near field capacity, was considered.

TABLE 2: Percentage of explained variance accounted for by clay content, C.E.C. and bulk volume on water content at different water potentials.

Water potential	Clay fraction < 2µm	C.E.C.	Bulk volume at field capacity
- 32 KPa	58	79	87
- 1.6 MPa	62	83	93
- 100 MPa	42	91	73

Water Regime and Soil Structure

Shrinkage is a very important aspect of Vertisols and related soils. Examination of volume changes, as a function of water loss on centimetric or larger clods, revealed a normal shrinkage in a large range of water potentials. This means that each volume of water lost is equivalent to the observed volume change. Consequently, a network of cracks is not normally observed on centimetric clods, and only pores resulting from clay particle arrangement become closer (Wilding and Tessier, 1988). Thus, water is located in very fine pores (< 1µm in size - Tessier, 1984), hydraulic conductivity of the system is low, and cohesion can become very high.

Consequently, in vertisols developed under very contrasted climate such as Vertisols in French Guadeloupe (Jaillard and Cabidoche, 1984), water distribution in the profile remains heterogeneous even after a long wet season. In France, this heterogeneity in water distribution occurs in summer but disappears during winter.

Conclusion

Even if the extension of vertisols in France is limited, numerous related soils with B vertic horizons are present. Except for studies carried out on Vertisols of Guadeloupe, research works on this subject were devoted to understanding vertic B horizons (Tessier et al., 1989-2). They also incorporate the main clayey soils developed in France. The soil structure developed under natural conditions during soil formation and weathering appeared as a determining factor to explain and predict soil behaviors. When possible, soil behavior has been studied in the laboratory on undisturbed and undried samples. There is no contradiction with classical studies using standardized data (tests).

Nevertheless, it is stressed that sample preparation can modify the system and only express a potential behavior, which does not

necessarily correspond to field behavior. Studying undisturbed sample changes by measuring water content, shrink-swell phenomena, and presence of air in soils is also a way to develop numerically express parameters which can be introduced in predicting models.

Literature Cited

Arrouays D. 1987. Carte des sols 1/50 000. Notice de la feuille de Bellegarde du Loiret. INRA-SESCPF, INRA publ. Versailles, 152 p.

Begon J.C. and Jamagne M. 1973. Les notions de pseudogley et de pelosol dans la signification des sols de la R.F.A. Science du sol 4:223-239.

Bonfils P. 1988. Carte pedologique de France 1/100 000. Feuille de Lodeve. INRA-SESCPR, INRA publ. Versailles.

Bonneau M., Duchaufour Ph., Le Talon F. and N'guyen Kha. 1967. Note sur quelques sols developpes sur substratum argileus. Pedologie XVII-1:106-108.

Bonneau M., Duchaufour Ph., Millot G. and Paquet H. 1965. Note sur certains sols vertisoliques en climat tempere. Bull. Soc. Geo. Als. Lorr. 17-4:325-334.

Bornand M., Dejou J. and Servant J. 1975. Les Terres Noires de Limagne: Leurs differents facies et leur place dans la classification francaise des sols. C.R. Acad. SC. Paris, 281-D:1689-1692.

Bruand A., Tessier D., and Baize D. 1988. Contribution a l'etude des proprietes de retention d'eau des sols argileux: importance de la phase argileuse. C.R. Acad. Ac. Paris. 307-II:1937-1941.

Bruand A. and Prost R. 1987. Effect of water content on the fabric of a soil material: an experimental approach. J. Soil Sci., 38:461-472.

Cabidoche Y.M. and Voltz M. 1987. Variations isotropes de volume en sols argileux heterogenes. Controle experimental dans le cas d'un vertisol calcique de Guadeloupe. p. 143-159. In Tranfert dans les milieux poreux deformables, INRA, Department de Science du sol, Versailles publ., p. 143-159.

Duchaufour Ph. 1982. Pedology. Pedogenesis and classification. George Allen and Uniwin Publ., London, 448 p.

Fedoroff N. and Fies J.C. 1968. Les vertisols du sud-est de la Beauce. Bulletin de l'Association francaise pour l'Etude du sol, 1:19-32.

Horemans P. 1984. Notice explicative de la carte pedologique de France a 1/250 000. Feuille de Paris. SESCPF, INRA publ., Versailles. 202 p.

Isambert M. 1984. Carte pedologique de France 1/100 000. Feuille de Chateaudun. Notice explicative. INRA-SESCPF, INRA, publ., Versilles, 259 p.

Jaillard B. and Cabidoche Y.M. 1984. Etude de la dynamique de l'eau dans un sol argileux gonflant: dynamique hydrique. Science du sol 3:187-198.

Jamagne M., Bliet L. and Remy J.C. 1970. Contribution a l'etude pedologique et agronomique des sols argileux du Bassin Parisien. La haute Brie. Annales Agronomiques 21-2:119-157.

Nguyen Kha. 1973. Recherches sur l'evolution des sols a texture argileuse en conditions temperees et tropicales. Thesis Univ. Nancy. 166 p.

Robert M., Hardy M., and Elsass F. 1989. Crystallochemistry and organization of soil clays derived from major desimentary rocks in France. Abstract Int. Clay Conference, Strasbourg, France, 1989.

Robert M., Isambert M., and Tessier D. 1973. Etude et premiees interpretations de l'evolution des glauconites dans les sols. C.R. Acad. Sc. Paris 277-D:1129-1139.

Salin R. 1983. Etude pedologique dans les Marais de Rochefort au 1/25 000. SESCPR-INRA, INRA publ., Versailles 200 p.

Tessier D. 1984. Etude experimentale de l'organisation des materiaux argileux. Thesis Univ. Paris VII INRA, Versailles publ. 361 p.

Tessier D. 1987. Identification of clays. Data from investigations with strongly hydrated systems. Proc. 20th Colloquium Int. Potash Institute, 1987, Int. Potash Int. publ., Bern, Switzerland 45-63.

Tessier D., Beaumont A., and Pedro G. 1990. Influence of rewetting rate on clay microstructure. Proc. Int. Micromorphology Meeting, San Antonio, Texas 1988, L. Douglas publ., 115-121.

Tessier D. and Berrier J. 1979. Utilisation de la microscopie electronique a balayage dans l'etude des sols. Observation de sols humides a differents pF. Science du Sol 1:67-82.

Tessier D., Bruand A., and Beaumont A. 1989. Relationship between clay minerzlogy and soil behavior in Paris Basin clayey soils. Abstract Int. Clay Conf. Strasbourg 1989.

Wilding L.P. and Tessier D. 1988. Vertisols: Their distribution, properties, classification, and management. Article in the book: Genesis fo vertisols: shrink-swell phenomena, 55-81, L.P. Wilding and R. Puentes ed., Texas A&M Univ. Publ., College Station, Texas.

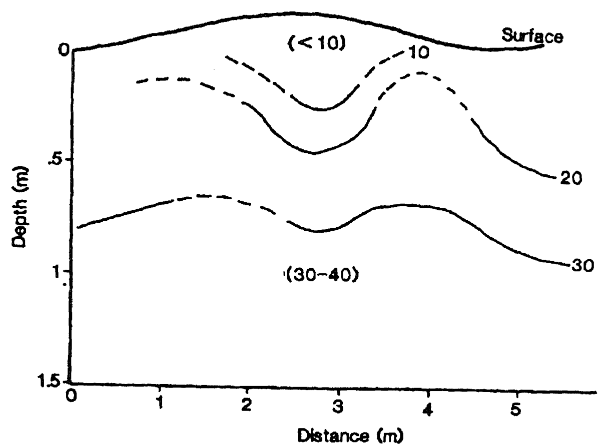


Figure 3: Minimum volumetric seasonal soil moisture content (%) (after Spotts, 1974).

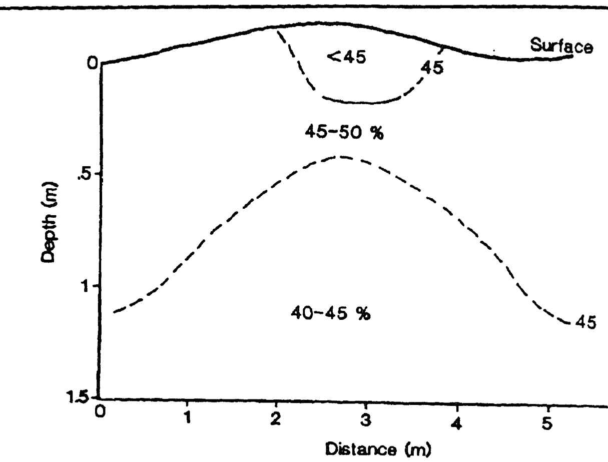


Figure 4: Maximum volumetric seasonal soil moisture content (%) (after Spotts, 1974).

bility of cracked, dry microlows is initially greater than that of microhighs; most water is transmitted down macrovoids to junctures with deep subsoil clay. Following recharge, the permeability of microlows becomes extremely slow. When wet, the permeability of Vertisol microhighs is reported to be 3-4 times greater than microlows. They found no consistent pattern for water retention relationships (i.e. field capacity, wilting point, or available water) between gilgai elements.

Spotts (1974) also observed that strain (percent dimensional change - either expansion or contraction) was in the zone of maximum change in soil water content (Fig. 6). The most active site was 45 cm below the surface of the microhigh. In the depression, the zone from 65-80 cm was the most active layer, but it was not as active as in the microhighs.

While these zones may express the greatest active movement, they may not correspond to zones which have the greatest shear failure, commonly observed between 1-2 m below the surface (Yaalon and Kalmar, 1978). Wilding and Tessier (1988) and Thompson and Beckmann (1982) postulated that the microlows would have the greatest change in moisture content between wet/dry states and greatest shear failure because of topographic position, cracking patterns and vegetative communities that extract water to greater depths and lower matric potentials. However, this may not be the case, based on Spotts (1974) work. Additional studies of this nature are needed to clarify soil moisture/strain relationships as a function of gilgai elements.

Table 2. General trends in selected physical properties between microlow and microhigh gilgai elements.

Property	Microlow		Microhigh
Cracking Expression (Width & Depth)	Higher	●●●>*	Lower
Water Movement			
External	Runon	<●●●	Runoff
Internal	Higher	●●●> <●●●**	Lower
Moisture Content (%)	Higher	●●●>	Lower
Clay Content (%)	Higher	●●●>	Lower
Specific Surface Area (m^2/g)	Higher	●●●>	Lower
COLE (cm/cm)	Higher	●●●>	Lower
Cohesive Strength	Lower	●●●>	Higher
Plasticity Index	Higher	●●●>	Lower
Shrinkage (%)	Higher	●●●>	Lower

*●●●> Commonly reported trends.
**<●●● Reverse trend has been reported.

Yule and Ritchie (1980) sampled microhigh and microlow gilgai elements of Houston Black (Udic Pellusterts) and Burleson (Udic Pellusterts) to compare soil shrinkage relationships. They observed higher 15-bar water retention and total shrinkage in microlows. Clay contents in microlows were also higher than in microhighs. The coefficient of linear extensibility (COLE) gave a good estimate of total vertical shrinkage but the expected 1:1 relationship was not found. They further observed that vertical shrinkage could be estimated from CEC almost as well as from COLE indices.

Expansion and contraction of Vertisols is related to the amount of water added or removed during seasonal cycles of desiccation/rewetting cycles (Wilding and Tessier, 1988). Measured elevational changes have been as much as 8 cm when dry soil is wet to field capacity (Aitchison and Holmes, 1953). Relative changes between gilgai elements are not well-known, but it is postulated by Thompson and Beckmann (1982) that, at different seasonal periods, the soil of one part of the gilgai complex may be relatively ex-

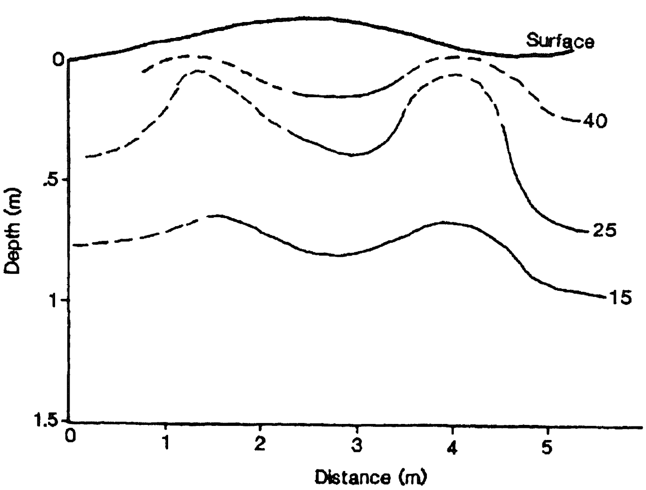


Figure 5: Maximum change in volumetric water content (%) from driest to wettest state (after Spotts, 1974).

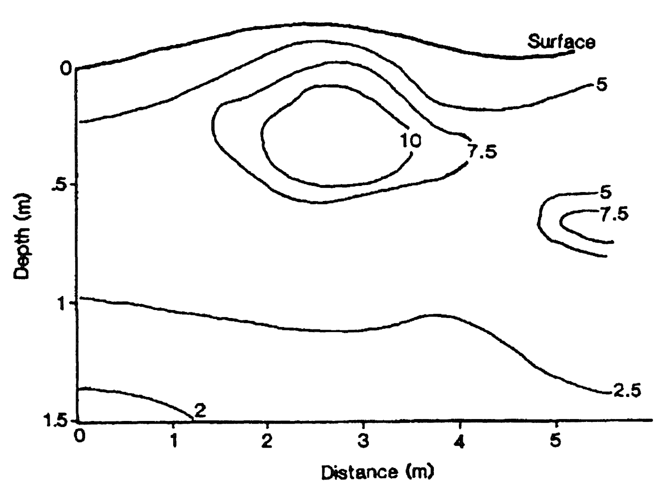


Figure 6: Percentage soil movement (strain) in microhigh and microlow (after Spotts, 1974).

panded, while the adjacent part is contracted, providing for maximal or minimal elevation differences.

For example, immediately following a wet period, the microhighs would begin to dry out even though water was ponded in adjacent microlows. However, when the water of the microlows has discharged, vegetation in these positions would flourish. Then, the microlows would dry out more rapidly than microhighs. With cracking, microlows would become desiccated to depths of 1 m or more while microhighs would still be moist at 20 cm below the surface (Thompson and Beckmann, 1982). Vegetation in microlows varies also with depth and duration of ponded water conditions; in deeper microlows, there is often no vegetation (Tucker et al., 1989). This is a fruitful field for further study that has been little explored.

Clay Contents and Clay Mineralogy

Thompson and Beckmann (1982) found no consistent pattern in clay content or clay depth functions between gilgai elements. Yule and Ritchie (1980) found that Vertisols in Texas have only slightly higher clay contents in microlows than in microhighs. These differences are likely in response to higher carbonate contents in microhighs. Likewise, clay mineralogy for Texas Vertisols was dominantly smectite with trace to small quantities of kaolinite, micas, and quartz for both microlows and microhighs. Calcite occurred in quantities reflecting total carbonate content in these soils. The layer clay minerals of these systems are indigenous to parent calcareous sediments (Dixon, 1982). Black Earths (mostly Pellusterts) of Australia have similar clay mineral suites (Thompson and Beckmann, 1982; Stace et al., 1968).

Cracking Patterns

Only sparse information is available on cracking patterns between microhigh and microlow gilgai elements. Yaalon and Kalmar (1978) recorded seasonal cracking and crack infilling trends in Vertisols of Israel, but these did not have gilgai. Thompson and Beckmann (1982) observed that a network of cracks develop mainly in the depressions during prolonged drying periods. They form rough circular patterns delineating mounds. Visible cracks on the mounds were less frequent and were apparently much finer and shallower than in depressions. In a comparison of paired gilgai elements of Black earths, Stirk (1954) found that shrinkage cracks developed at lower moisture tensions in depressions than in mounds. Verification of cracking patterns, crack closure hysteresis, and cracking depths as a function of seasonal soil moisture within gilgai elements is an acute area of research need.

Chemical Properties

Table 3 provides some generalized relationships between selected chemical properties of the microhigh and microlow gilgai elements. While general trends are evident for many properties, reversals are apparent for some of the more temporal conditions influenced by seasonal sampling period and site specificity (i.e. soluble salts, ESP, etc.).

requires a multi-discipline approach including: expertise of a field soil scientist to locate a representative sampling site; local soil conservation staff to obtain permission for access to property and to ascertain past cultural history; a plant taxonomist to make a detailed vegetative survey of gilgai elements; an engineer to conduct a detailed topographic survey of the specific sites; soil scientists to describe and sample the selected site; and laboratory analysts to characterize the physical, chemical, and mineralogical properties.

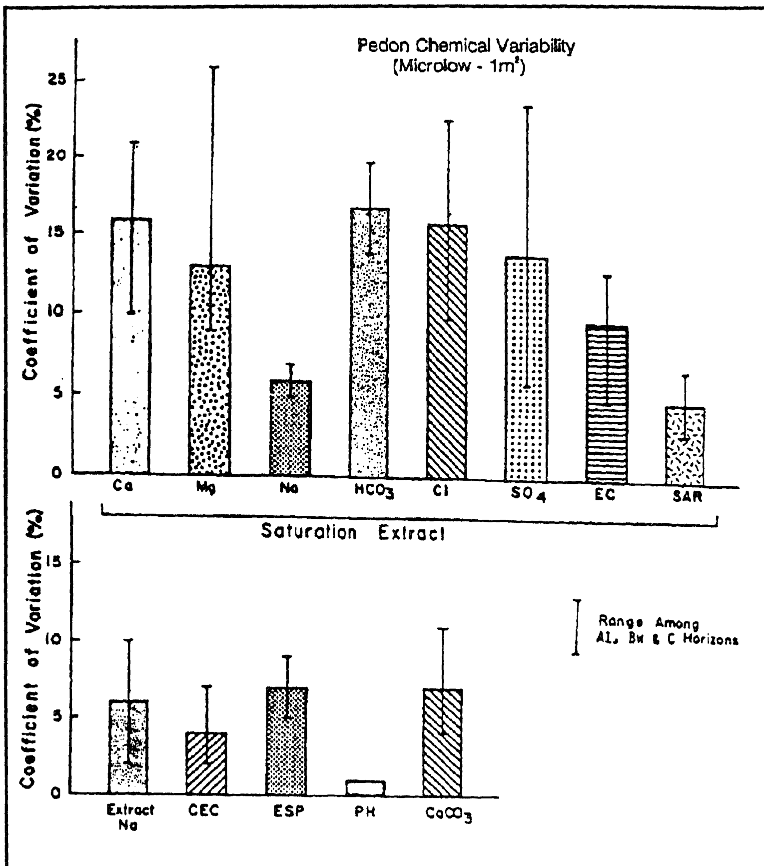


Figure 8: Observed chemical variability within a sampling unit of 1 m² for Houston Black (Udic Pellustert) microlow gilgai element (after Wilding, 1985).

Selection of a Representative Sampling Site

Perhaps one of the most critical steps in a microvariability study is making sure the site sampled is a good representation of the field conditions for a given soil. This phase of the work was conducted by co-author, Mr. Wesley Miller, Area Soil Scientist, USDA-SCS, Victoria, Texas, who was party chief for the soil survey of Victoria County.

To initiate site selection, six alternative areas from numerous mapping units of the Lake Charles clay within Victoria County were considered. Several conditions were imposed on the selected areas, namely, they had to be: 1) within large mapping units of the Lake Charles clay; 2) uncultivated; and 3) have gilgai cycles with periodicities between 5-15 m. Periodicity of gilgai and prospective sites were screened from the office by viewing older aerial photosheets of the survey area (1:14,840 and 1:7,920 scales). Ground truth was collected from near the center of the mapping units with linear point transects at 0.6 m intervals such that at least three consecutive microhighs and microlows were crossed. Field transects were oriented with calcareous microhighs in the middle of the transect, and the transect run in a straight line or at a slight angle to an adjacent microhigh.

At each of the transect points, a 4 cm core was taken with an hydraulic power probe to at least 1 m depth (a few to 1.75 m) and a soil description written. Table 4 illustrates gilgai periodicity, diapir periodicity, length of transect with surface colors of specified thickness, and pedon classifications for four of the six transects run. Two transects were not included because they inadvertently were oriented lengthwise along microhighs, which inflated the percentage of Entic versus Typic Pellusterts observed.

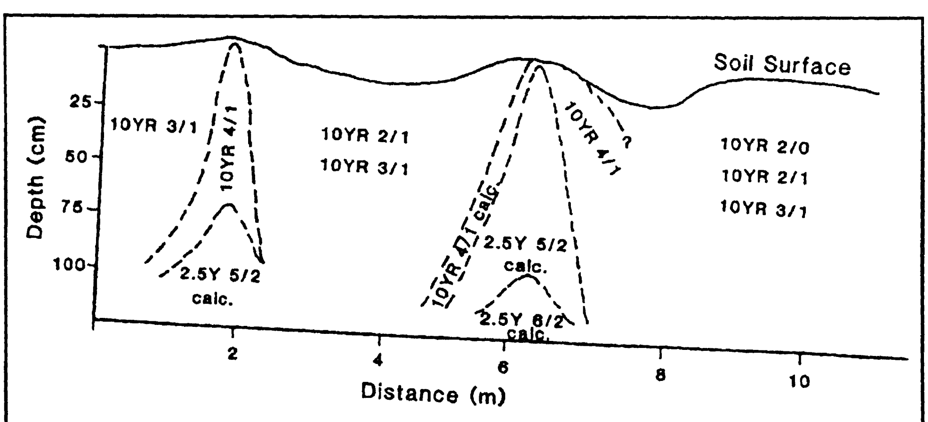


Figure 9: Elevation and soil transect of Lake Charles clay (Typic Pelludert).

In summary, the mean periodicity of the gilgai cycles

was 4.9 m and ranged from 3.1-7.0 m. Local relief differential between adjacent microhighs and microlows ranged from 15-35 cm and centered on 20 cm. By pedon classification, 71% of the 24 pedons examined were Typic Pelluderts (≥ 30 cm of 10YR 3/1 or darker surface horizons over half of the pedon) while 29% were Entic Pelluderts (colors not meeting Typic and most being 10 YR 4/1 or lighter). Considering the proportion of the 58.5 m total transect distance, 65% of the linear distance consisted of Typic Pelluderts and 35% Entic Pelluderts. Figure 9 illustrates a crossectional profile of one of these transects very similar to the site selected later for detailed study.

Table 4. Gilgai periodicity, diapir periodicity, length of transect with given surface thickness/color criteria and pedon classification for Lake Charles clay, Victoria County, Texas.

Transect #	Peridicity (m)		Pedon No.**	Pedon Length		Pedon Length with Surface Colors		Pedon Classification***	
	gilgai	diapir*		gilgai	diapir**	10YR 3/1 or darker (≥30 cm thick)	10YR 3/1 (<30 cm thick) & 10YR 4/1 or lighter	Typic Pelluderts	Entic Pelluderts
5	6.30	6.10	1	3.15	3.05	2.44	0.61	+	
			2		——	1.83	1.22	+	
	9.70	6.10	3	4.85	3.05	2.44	0.61	+	
			4		——	1.83	1.22	+	
		3.66	5		1.83				
			6		——	0.00	1.83		+
6	6.10	5.80	1	3.05	2.90	1.37	1.52		+
			2		——	1.98	0.91		+
	4.25	4.26	3	2.12	2.13	1.52	0.61	+	
			4		——	2.13	0.00	+	
	4.85	4.26	5	2.43	2.13	2.13	0.00	+	
			6		——	1.22	0.91	+	
7	6.10	7.00	1	3.05	3.50	1.98	1.52	+	
			2		——	1.83	1.68	+	
	6.00	5.80	3	3.00	2.90	1.22	1.68		+
			4		——	2.90	0.00	+	
	4.00	3.04	5	2.00	1.52	1.52	0.00	+	
			6		——	1.52	0.00	+	
8	4.26	4.58	1	2.13	2.29	1.68	0.61	+	
			2		——	1.83	0.46	+	
	3.35	3.04	3	1.68	1.52	0.61	0.91		+
			4		——	0.30	1.22		+
	3.35	4.88	5	1.98	2.44	1.53	0.91	+	
			6		——	1.37	1.07	+	
Total		58.52	24			37.94	20.57	17	7
· Typic Pelluderts****						65%		71%	
· Entic Pelluderts							35%		29%

*Diapir defined as zone with thinnest and lightest color portion of microhigh
**Pedon defined as 1/2 distance between diapirs
***Pedon classification based on 1/2 diapir periodicity as concept of pedon
****Pedon classification based on total length of transects with given surface colors of specified thickness

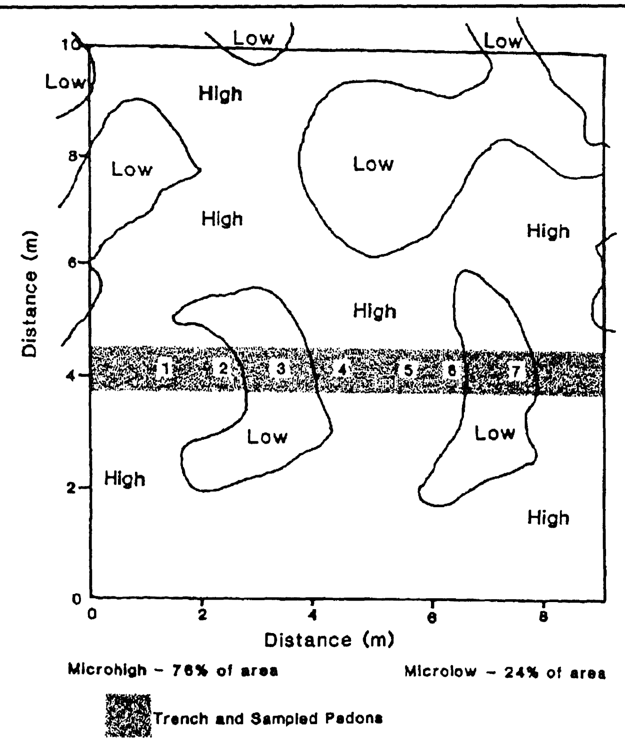


Figure 10: Microtopography, trench and sampled pedons in Lake Charles clay (Typic Pelludert).

Several insights about the short-range spatial variability gained from these detailed mapping unit transects were as follows:

1. Calcareous diapirs (chimneys) were more pronounced and occupied a larger percentage of the smaller diameter microhighs than the larger diameter microhighs (20-35%).
2. The pedon is different depending on whether it is defined as one-half of the gilgai periodicity or one-half the distance between the thinnest and lightest-colored surface horizons (diapir periodicity). This definition difference will influence the pedon concept and classification. A higher percentage of Entic Pelluderts will occur if the pedon is based on one-half the gilgai periodicity rather than one-half the diapir periodicity.

3. Calcareous diapirs are not always in the center of the microhigh but may be on the shoulder or intermediate position between a microhigh and microlow. Sometimes, they are a small, positively elevated area in the middle of the microhigh. They are most easily identified by 2-5% calcium carbonate nodules on the surface and associated gray or grayish-brown clay at the surface or within 2-6 cm of the surface.
4. The transition area (10YR 4/1 colors) between microhigh chimneys and microlows occupies more area than previously considered.
5. Microhighs are more variable in morphological properties than microlows. Better classification and morphological data could be obtained if a 10 m² area were gridded at 0.5-1 m intervals with elevation control to yield a 3-D topographic surface net of topography and horizonation.

Table 5. Vegetative composition of microhighs and microlows for Lake Charles clay at ICOMAQ site in Victoria County, Texas.

Species		Composition (%)			
		Microhigh		Microlow	
Common Name	Scientific Name	Mean	Range	Mean	Range
Grasses (perennial):					
Indiangrass	Sorghastrum nutans	—	—	—	—
Brownseed paspalum	Paspalum plicatulum	1.5	0.7-2.8	.4	2.4-8.6
Unknown paspalum (#10)	Paspalum spp.	—	—	0.1	0.0-0.5
Texas wintergrass	Stipa leucotricha	4.6	2.8-8.1	3.7	1.0-3.7
Little bluestem	Andropogan saccharoides	—	—	—	—
Broomsedge bluestem	Andropogan virginicus	3.6	2.8-4.2	1.3	0.0-3.7
Bushy bluestem	Andropogan glomeratus	—	—	0.1	0.0-0.4
Scribner's dicanthelium	Panicum scribnerianum	0.3	0.0-1.1	—	—
Dropseed species (smutgrass)	Sporobolus spp.	3.7	2.7-4.2	1.3	0.0-20.0
Blue-eyed grass species	Sisyrinchium spp.	2.2	1.1-2.4	0.1	0.0-0.4
Little barley	Hordeum pusillum (Nutt.)	—	—	0.7	0.5-1.6
Fescue	Festuca spp.	—	—	0.1	0.0-0.4
Phalaris species	Phalaris spp.	—	—	1.0	0.5-2.4
Common bermudagrass	Cynodon spp.	—	—	0.6	0.0-1.9
Unknown grass #8		—	—	—	—
		15.9		**13.4**	
Forbs (perennial):					
Yellow nutsedge	Cyperrus esculentus	8.5	3.8-13.6	0.9	0.0-2.4
Eleocharis species	Eleocharis spp.	—	—	8.5	6.3-13.4
Yellow woodsorrel	Oxalis dillenii	2.3	1.0-6.4	3.5	2.8-4.3
Upright prairie coneflower	Rudbeckia spp.	0.9	0.0-1.6	—	—
Dotted gayfeather	Liatris punctata	—	—	—	—
Button snakeroot	Liatris pycnostachya	—	—	—	—
Thistle species	Cirsium spp.	—	—	—	—
Maximilion sunflower	Helianthus maximiliani	—	—	—	—
Houstonia species	Houstonia spp.	—	—	—	—
Indigo species	Baptisia spp.	1.7	0.0-5.9	—	—
Yellow neptunia	Neptunia lutea	0.6	0.7-1.1	0.9	0.0-1.9
Butterfly pea	Centrosema virginianum	—	—	0.1	0.0-0.4
Catclaw sensitivebriar	Mimosa strigillosa	1.9	0.3-4.5	—	—
Plantain species	Plantago spp.	0.9	0.0-2.2	0.3	0.0-0.5
False dandelion	Pyrrhopappus spp.	2.5	2.5-3.4	0.3	0.0-1.0
Frogfruit	Phyla incisa	—	—	1.8	0.0-3.1
False garlic	Nothoscordum spp.	0.3	0.0-1.1	0.4	0.0-0.8
Goldenrod	Solidago spp.	4.0	1.7-8.0	4.7	5.3-7.3
Green antelopehorn	Asclepias virdiflora	—	0.0-0.3	0.4	0.0-1.3
Western ragweed	Ambrosia psilostachya	0.6	0.0-1.41	2.4	2.0-2.9
Euphorbia spp.	Euphorbia spp.	4.2	3.4-4.3	8.1	0.0-13.4
Eryngo species	Eryngium spp.	—	—	1.4	0.5-2.8
Unknown forb #4		0.3	0.0-0.7	0.7	0.0-1.6
Unknown forb #6		2.8	0.0-5.6	2.8	0.0-5.3
Unknown forb #7		1.1		1.3	0.0-3.2
		32.6		**38.5**	
Forbs (annual):					
Annual broomweed	Xanthocephalum sphaerocephalum or Xanthocephalum texanum	0.3	0.0-0.7	1.3	1.4-2.4
Unknown annual forb #9		—	—	—	—
Unknown annual forb #10		—	—	—	—
Miscellaneous forbs		0.5	0.0-1.1	0.3	1.0-2.8
0.8		**1.6**			
Bare Ground		38.6	20.0-50.0	14.4	5.0-25.0
Litter		12.1	5.0-25.0	32.1	20.0-50.0
		100%		**100%**	

After field transect studies were conducted, a site was selected for detailed study which reflected the mean conditions of transects, for which access could be gained and for which a pit could be opened and remain open for about one week. Transect #7 (Table 4) represents the area chosen for the field study and is close to where the trench was placed in the detailed study area. It is also close to the mapping unit type location of Lake Charles clay in the Victoria County Soil Survey.

Detailed Study Area

After selecting the representative study area, three transects were run across multiple microhighs and microlows to determine trench location and orientation for best light exposure. To locate calcareous diapirs in the center of microhighs, the trench face was "V" shaped with the apex in about the center of the pit wall. At that point, the pit was about 10° off the straight line.

Before the trench was excavated, the location of microhighs and microlows were sketched to scale and the location of the trench placed on this sketch (Fig. 10). The pit extended through

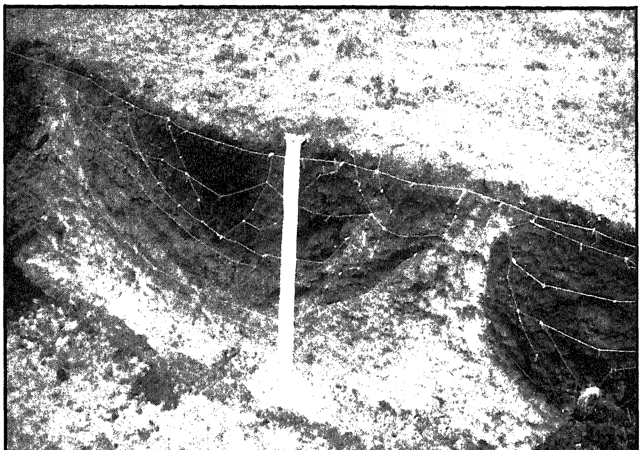


Figure 11: Distinct soil zones or polyhedrons are delineated by white string.

two microhighs and two microlows. The study area was approximately 10 m², but varied, depending on size and shape of gilgai. Study area boundaries were outlined for detailed vegetative and topographic surveys.

Vegetative Survey

The species composition, bare ground, and litter were recorded for three microhighs and three microlows. Compositional percentages (means and ranges) for the plant survey are given in Table 5. These data were collected at the site of the ICOMAQ study, within a few kilometers of the detailed study area. The range condition of the study area was severely overgrazed. The percentage bare ground is higher and plant litter lower on microhighs than microlows, reflecting areal percentage of apparently active calcareous diapirs that rise to the surface of microhighs.

Species preference between microhigh and microlow positions was not strong but noted differences were as follows: Brown seed paspalum (microlow), broomsedge bluestem (microhigh), rattail smutgrass (microhigh), dotted blue-eyed grass (microhigh), yellow nutsedge (microhigh), Eleocharis spp. (microlow), false dandelion (microhigh), western ragweed (microlow), and Euphorbia marginhea (microlow). The lack of a more striking difference in species composition between gilgai elements may be due to the fact that this site was overgrazed. It also may reflect the seasonality of the vegetative survey with dynamic changes in vegetative composition occurring with changing moisture contents.

Topographic Survey

Within the 10 m² study area, a topographic survey is under construction on a grid at 1 m intervals; recordings will be taken at closer intervals as needed. This work is still in progress. Future studies will include surface cracking patterns as a function of soil moisture conditions changing with time that will be drawn to scale with reference to microtopographic highs and lows of the study area.

Trench Description and Sampling Scheme

A pit was excavated along the transect selected to cross two microhighs and two microlows. The major pit wall was slightly "V" shaped with the trench face 1.2 m wide, 2 m deep and 10 m long. A "T" leg about 1/3 the length of the pit face was constructed in a wedge-shaped design to allow better light exposure of the pit face during midday. The "T" extended from the bottom of the pit and sloped upward to the surface over a length of about 5 m for easy access.

Following excavation, the entire pit face was picked to expose horizon configuration, natural structural peds, and slickenside orientation. After careful examination of all observable features, nails were inserted into the face to outline all areas having common features such as color, structural arrangement, slickenside patterns, carbonate and ferromanganese segregations, and other distribution patterns. White string was used to connect the nails and enclose common distinct soil zones or polyhedrons (Fig. 11).

Seven pedons (sampling units) were selected for detailed sampling, which included two microlows, two microhighs, and three intermediate positions. The pedons were sampled for field moisture content, shortly after pit excavation, at depths of 10, 60, 110 and 190 cm. This was to establish initial moisture conditions relative to marked differences observed in moist soil consistencies between gilgai elements (Table 6). The Bk horizons of microhighs had much firmer soil consistency than microlows, and it was hypothe-

Table 6. Spatial variability in soil moisture content (weight percentage) of Lake Charles clay

Depth (cm)	Lateral site location						
	Pedon #1 (MH)*	Pedon #2 (Int.)	Pedon #3 (ML)	Pedon #4 (Int.)	Pedon #5 (MH)	Pedon #6 (Int.)	Pedon #7 (ML)
10	26.8	27.1	28.7	25.6	23.6	28.6	32.0
60	23.9	22.4	25.5	24.9	23.0	24.6	23.8
110	18.5	18.3	20.9	23.3	22.6	21.5	25.9
190	23.2	20.0	19.8	23.7	23.0	22.8	21.9

* Gilgai position - MH - microhigh
Int. - intermediate between MH and ML
ML - microlow

sized that this may be due to different moisture states. While microlows did tend to have slightly higher moisture contents than microhighs or intermediate positions, differences were not great nor consistent. At 190 cm, microhighs had slightly higher moisture retention than microlows. Differences in moisture content observed do not seem to be adequate to explain the distinctly greater cohesive strength and firmness of microhigh subsurface structural units vs. microlow units.

Detailed profile descriptions were written of the seven pedons. Bulk samples and bulk density clod samples were taken of each horizon in all pedons. Bulk samples of distinct soil zones (layers) that were not collected in the seven pedons sampled were collected as satellite samples. All samples were shipped to the National Soil Survey Laboratory, Lincoln, Nebraska, for laboratory characterization analyses, except for field soil moisture contents, which were determined in the Texas Soil Characterization Laboratory, Texas A&M University, Soil and Crop Sciences Department, College Station. Laboratory data are not yet available except for initial moisture status.

Selected morphological characteristics along the pit face were plotted on graph paper in the field at a scale of 1:4. The properties were plotted from a string line placed on the level about 25 cm above the microhighs for elevation control (Fig. 11). From this crossectional profile, the percentage of the pit face comprised of microhigh, intermediate and microlow gilgai elements, with corresponding horizon percentages of given colors, was planimetered or determined from grid areas (Table 7). These data will be discussed under the next section. The crossectional profile data also were input into a Geographical Information System (GIS, ARCINFO) at South Technical Center, Fort Worth, Texas, so morphological and analytical databases may be overlayed. Further, this allowed presentation of the crossectional profile at variable scales. GIS reductions the crossectional profile and corresponding areas with similar zip patterns are presented in Figures 12 and 13.

Table 7. Percentage of pit face comprised of microhigh, intermediate and microlow gilgai elements with corresponding horizon percentages of given colors.

Gilgai element	Gilgai element (%)	Horizonation (%)	Horizonation Kind	Soil colors (Munsell Notations)
Microlow	29	72	(A, A2, Bw)	10YR 2/1, 3/1
		8	(Bk)	10YR 4/1 to 6/1
		20	(Bk)	5Y and 2.5Y 6/2, 7/2
Intermediate	52	9	(A)	10YR 2/1, 3/1, 4/1
		49	(Bk)	10YR 4/1, 5/1 or 6/1
		42	(Bk)	5Y and 2.5Y 6/2, 7/2
Microhigh	19	4	(A)	10YR 4/1
		18	(Bk)	10YR 5/1
		78	(Bk)	5Y and 2.5Y 6/2, 7/2
Total Trench		38	(A, A2, Bw, Bk)	10YR 2/1, 3/1
		8	(A, Bk)	10YR 4/1
		22	(Bk)	10YR 5/1, 6/1
		32	(Bk)	2.5Y and 5Y 6/2, 7/2

Summary of Field Observations of Sampling Site

The site selected was an excellent example of the extreme variability in many Vertisol properties corresponding to microhighs and microlows. We hope this study will illustrate the complexity of soil morphology, chemical, physical, and mineralogical properties, water movement, root

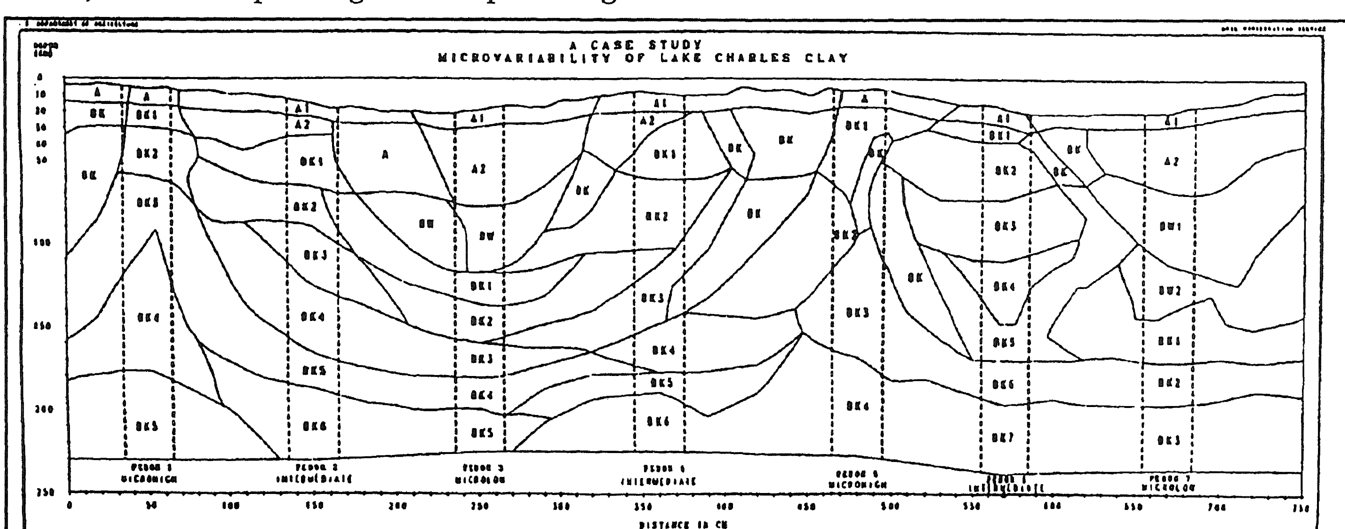


Figure 12: A reduction of the field-constructed crossectional profile using a Geographical Information System (ARCINFO).

development, and crop/vegetative adaptivity and response to microlow and microhigh gilgai topographic elements. Following are a few specific notes and suggestions relative to this case study:

Morphological Properties Associated With Gilgai Elements

Striking morphological differences were observed in three gilgai positions, namely, microhigh, microlow, and intermediate positions. The two microhighs of pedons #1 and #5 had were narrow (30-70 cm wide) diapirs "chimneys" of grayish, calcareous clays extending from the lower Bk horizon to the surface. They did not appear to be due to crack infillings, but rather, these "tepee-shaped" structures appear to have been pushed or squeezed up cat steep angles along slickenside planes that border microlows. The chimneys are not always located in the center of the microhighs but may be offset onto the intermediate position.

The lows are generally black clays with little or no carbonates either as nodules, soft segregations, or disseminated carbonates. In lower Bw and Bk horizons, a few carbonate nodules occur randomly distributed throughout these horizons but lack soft carbonate coatings or encasements that are common in Bk horizons of microhigh and intermediate positions. It appears that these carbonates are undergoing dissolution and leaching because they are rough-surfaced, entirely free of soft segregation coatings, and exhibit clean sand grains protruding from carbonate nodules.

Between the microlows and microhighs was an extensive zone of calcareous, gray clay, darker in color than the microhigh, but lighter than the microlow. It contained common carbonates as soft segregations, hard concretions encased in soft calcareous rinds, and disseminated forms. These calcareous materials were commonly banded at angles corresponding to bordering slickenside planes and oriented towards the microhigh. Further, they were virtually continuous with lower Bk horizons.

In the microlow of pedon #5, the Bw and Bk horizons had circular crossectional, pedotubule - structures about 3-5 cm in diameter. They were oriented vertically and are quite possible krotovinas from crayfish activity.

Quantification of Gilgai Elements

Quantitative determinations of the crossectional area of the trench face comprised of different gilgai elements and corresponding horizonation are given in Table 7. Over half (52%) of the trench was comprised of the intermediate gilgai position, followed by the microlow (29%) and the microhigh (19%). Black or very dark gray A and Bw horizons comprised nearly three-fourths of the microlow, with 28% comprised of lighter gray Bk horizons.

In contrast, intermediate and microlow positions were dominated by gray and grayish-brown Bk horizons (91% and 96%, respectively), with the remainder comprised of thin black, very dark gray, or dark gray A horizons. Of the total trench, 38% was comprised of black or very dark gray clays, 8% by dark gray clays, 22% by gray clays, and 32% by light gray and light, brownish-gray calcareous clays.

Slickenside Orientation and Spacing

Large slickenside planes tend to outline the microlow and microhigh gilgai elements; they

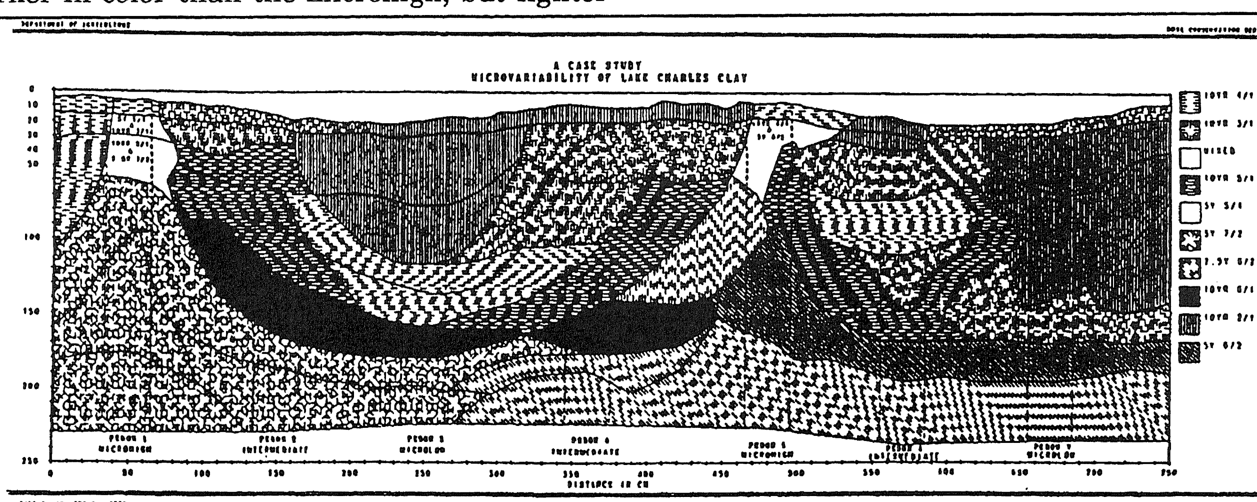


Figure 13: Crossectional profile illustrating zones within trench with common soil colors. Graph drawn using a Geographical Information System (ARCINFO).

form a "bowl structure" or "cone of revolution" in the microlow where the angle of dip (plane of the pit face) at the apex or base was only about 15°. Along the upward-tending diapirs of the microhighs, the angle of dip was steep (60-75°). In the trench exposure, there were 6 or 8 large slickenside planes which were carefully hand-picked back into the face of the pit (Fig. 11). These planes were 2-5 cm apart as they approached the diapirs of the microhigh but 4-10 cm apart in the intermediate and microlow pedons. The major slickenside faces were polished and grooved with a network of ridges and valleys. The ridges had a local relief differential of 1-6 cm higher than the troughs. The troughs were 4-25 cm wide. Roots tended to follow slickenside faces but some extend through the faces.

Starting at a microhigh of pedon #1, the slickenside planes extended downward at a dip angle of 60-70° from the horizontal. The slickenside planes ran in a diagonal concave curved arc across the intermediate pedon #2, to 180 cm depth, where they flattened out to 15-20° in the microlow of pedon #3. The same set of major slickensides extended in a concave curved arc upward across the intermediate pedon #4, to 63 cm below the surface of the microhigh of pedon #5, where the angle of inclination was 60-75°. Similar slickenside planes were observed between intermediate pedon #6 and microlow pedon #7.

Perpendicular to the pit face, the angle of strike for major slickensides ranged from 25-60° but was commonly 50-55°. No pattern for strike angle was noted relative to angles of dip or gilgai positions.

Carbonate Forms and Patterns

As previously mentioned, common to many, soft and hard segregations of carbonates are randomly distributed in lower Bk horizons of all pedons sampled and ubiquitous to all Bk horizons of intermediate and microhigh positions.

In pedon #1, a zone of soft segregations extended to within 53 cm of the surface and followed a downward concave arc along slickenside planes towards the microlow of pedon #3. In pedon #5, soft segregations were translated along slickenside planes to near the surface on the extreme right side of this pedon and follow the edge of a large slickenside plane between pedons #5 and #6.

The microlows of pedons #3 and #7 are either leached of carbonates or have a few hard carbonate nodules but no soft segregations of carbonates. Along the outer concave "bowl" edges of the microlows, hard carbonate concretions occur in a 20-30 cm wide band in the shape of a concave upward arc. This is believed to have been translated upward from the lower Bk between two bordering slickenside planes. Adjacent to this zone of hard concretions is a 10-15 cm wide band of both hard concretions and soft segregations of carbonates.

Distribution of Fe-Mn Nodules

Few to common, fine, distinct ferromanganese rounded nodules and segregations occur throughout the Bk horizons of all gilgai elements. In the microlow, they also occur in lower Bw and Bk horizons. The nodules are most pronounced in the strongly calcareous lower Bk horizons. It is not clear whether these segregations represent contemporaneous alternating redox states or may be relict from wetter paleoenvironments.

Based on current measurements of hydrology and redox in a Lake Charles site for the ICOMAQ Study (Griffin et al., 1989), it does not seem reasonable that the Fe-Mn segregations are contemporaneous because, over the past year, neither the microhigh nor microlow has been saturated. Further, the redox states would need to be very low to reduce oxidized phases of these compounds in the presence of free carbonates. An alternative explanation is that most of the Fe-Mn segregations are the result of more strongly reducing conditions following post-deposition of the deltaic-fluvial Beaumont formation sediments. Such conditions would have been present during the period of marshy environment up through the dewatering stage and ripening of these sediments.

Future Aspects

Close-interval microvariability in Vertisols that corresponds to gilgai elements, oscillatory horizonation, and contorted or interrupted zones has received inappropriate attention. To better elucidate soil genesis, classification, management, use, and distribution of Vertisols, the following areas justify increased research emphasis: sampling schemes, hydrological behavior, cracking patterns, stress/strain/shear failure, root distribution patterns, soil chemistry/fertility properties, and plant distribution/response relationships. These databases should be assembled over a broad geographical area considering macro/micro-climate, parent material, vegetation, soil management, and land use variables. Such knowledge will serve as the

foundation for reevaluating the adequacy of the pedon concept for Vertisols, sampling schemes, engineering qualities, pollution hazards, cultural practices, and technology transfer under mechanized and subsistence agriculture.

Sampling Schemes

The sampling design used in this study is an approach to better represent and quantify the spatial distribution of properties. It suggests that positions intermediate between microhighs and microlows comprise a significant landscape component. In future studies, the trench should be cut perpendicular to the major face to allow better visualization and sampling of spatial variability in the third dimension, thus fostering a 3-D concept of spatial variability.

In field observations of core samples, the sampling scheme as a 1 m grid, with appropriate satellite observations at shorter intervals, would allow the most times and cost-effective method to quantify morphological variability. Such a grid should be accompanied by elevation control so computer-generated topographic surface nets can be constructed for horizons and features of interest. The grid should be of sufficient size to include a minimum of three microhighs and three microlows.

Proposed Revision in Pedon Concept

The concept of a pedon as a sampling unit should be reappraised and revised to encompass the variability within an area no greater than 1 m^2. In Vertisols, such an area will include significant variability in some cases and relatively little in others. Such a sampling scheme will focus attention on variable topographic and horizonation elements within the landscape system and not bias the data towards microlows. This concept would also bring sampling design into conformance with other soils with less variable attributes. Current state of soil behavior knowledge demonstrates the importance of verifying spatial variability in the physical, chemical, and biological properties of these soils over short lateral distances. The revised concept would further enhance such a microvariability knowledge base.

If such a pedon revision were accepted, then the mapping and classification of Vertisols would not be so dependent on the 7 m cycle of gilgai elements. The classification would be based on the distribution of attributes found within the mapping unit and or their pattern of occurrence. In many cases, a complex of two or three soil conditions would be recognized, based upon verified microvariability and the importance of these conditions to use and management. In many Vertisols, insufficient variability may occur within the polypedon area to justify a complex; in this case, the soil would be mapped as a consociation.

Such a revision also would be consistent with the definition of the pedon as a sampling unit and the polypedon as the classification unit with landform expression. The pedon would be standardized so it was of fixed dimensions and not variable in response to diapir or gilgai periodicities. The latter generates ambiguity in pedon definition and application for soil classification.

Hydrological Behavior

Water and chemical transfer through Vertisols has been studied extensively (Ritchie et al., 1972; Kissel et al., 1973; Bouma, 1988) but relatively little information is available to quantify soil moisture relations and transfer between gilgai topographic elements (Spotts, 1974). For example, what is the saturated and unsaturated hydraulic conductivity between these microsites, what influence do slickenside planes have on these hydraulic parameters, and what is the relative flux of water and chemical constituents between microlows and microhighs?

Monitoring saturated and near-saturated soil moisture conditions between microlows and microhighs has been initiated through an ICOMAQ Study in Texas (Griffen et al., 1989). Piezometers, tensiometers, and boreholes have been replicated in a series of microhigh and microlow sites of the Lake Charles clay (Typic Pelluderts).

Seasonal influences on runoff/recharge/discharge characteristics and evapotranspiration that may be influenced by cracking, rooting, and macrostructural properties are also of vital interest to plant distribution patterns and responses to hydrological behavior for irrigation and drainage (Bouma, 1988). Likewise, such dynamics influence nutrient availability, leaching losses, mineralization, reduction, and immobilization of plant nutrients (Kissel et al., 1973; Wilding and Rehage, 1985). Such knowledge has vast agronomic and engineering implications but to date is not available.

Soil Fertility Relationships

Plant response due to spatial variability in Vertisols is not well partitioned to physical, chemical, and biological effects or interactions.

In Texas, few accounts are available showing that crop response is variable from microhigh to microlow gilgai positions, though native vegetation frequently corresponds to these microsite environmental habitats. It is apparent from Australian research (Russell et al., 1967; Stace et al., 1968; Thompson and Beckmann, 1982) that crop responses are well correlated with differences in gilgai soil properties. It would appear that the interactive effect between soil moisture patterns and soil chemistry/fertility patterns should be further appraised.

In Texas, on Vertisols cropped to cotton, attempts were made to relate gilgai patterns to the incidence of cotton root rot (Phymatotrichum omnivorum), but this research did not verify such a correlation. In Vertisols of the Gulf Coast Prairie region of Texas, Fe chlorosis patterns appear to be related to gilgai spatial variability, but this hypothesis is untested.

Soil fertility relationships to gilgai offer another fruitful area for research in the U.S, especially considering the ability to apply differential rates of water, chemicals, herbicides, and pesticides with computer-controlled, real-time sensing capabilities.

Stress/Strain/Shear Failure Relationships

Until stress/strain/shear failure interactions are better verified as related to changes in soil moisture patterns of gilgai elements, a more comprehensive knowledge of Vertisol genesis, management, and use implications is precluded. For example, we do not have a verified database on extent of elevation changes, expression of slickenside failure zones, soil strength attributes, or cracking patterns between gilgai microsites. There are few studies of the 3-D shear failure structural analysis in unsaturated soil zones (Knight, 1980). These soil mechanical properties are just as germane to agronomic practices and crop management as to construction engineers. A better verification of these properties would enhance our knowledge of irrigation frequency, cracking dynamics, pattern and spacing of gilgai, soil tillage attributes, and nutrient use efficiency. Future studies of this nature should be initiated only after close coordination, liaison, and collaboration with civil and agricultural engineers with expertise in soil mechanics.

References

Ahmad, N., 1983. Vertisols. In L.P. Wilding, N.E. Smeck and G.F. Hall (eds.). Pedogenesis and Soil Taxonomy II. The Soil Orders. Developments in Soil Science IIB. Elsevier Publ. Co., Amsterdam, pp. 91-123.

Aitchison, G.D. and J.W. Holmes, 1953. Aspects of swelling in the soil profile. Aust. J. Appl. Sci., 4:244-259.

Beckmann, G.G., G.D. Hubble and C.H. Thompson, 1970. Gilgai forms, distribution and soil relationships in north-eastern Australia. In Proceedings of Symposium on Soils and Earth Structures in Arid Climates, Sydney, pp. 116-121.

Bouma, J. and J. Loveday, 1988. Characterizing soil water regimes in swelling clay soils. In Vertisols: Their Distribution, Properties, Classification and Management, Texas A&M University Press, College Station, Texas.

Bouma, J. and P.A.C. Raats, 1984. Proceedings of the ISSS Symposium on Water and Solute Movement in Heavy Clay Soils. ILRI Publication No. 37, Wageningen, The Netherlands.

Dixon, J.B., 1982. Mineralogy of Vertisols. In Vertisols and Rice Soils of the Tropics, Symposia Papers II, 12th ICSS, New Delhi, pp. 48-59.

Dudal, R. and H. Eswaran, 1988. Distribution, properties and classification of Vertisols. In L.P. Wilding and R. Puentes (eds.), Vertisols: Properties, Classification and Management. Texas A&M University Press, pp. 1.

Dudal, R., 1965. Dark clay soils of tropical and subtropical regions. FAO Agricultural Development Paper No. 83, Rome.

FAO, 1983. Proceedings of the 5th Meting of the Eastern African Subcommittee for Soil Correlation and Land Evaluation. Wad Medani, Sudan. World Soil Resource Report No. 56, Rome.

Hallsworth, E.G., 1968. The gilgai phenomenon. In H.C.T. Stace, G.D. Hubble, R. Brewer, K.H. Northcote, J.R. Sleeman, M.J. Mulcahy and E.G. Hallsworth (eds.). A Handbook of Australian Soils. Rellim Tech. Publ., Glenside, South Australia.

Hallsworth, E.G., G.K. Robertson and F.R. Gibbons, 1955. Studies in pedogenesis in New South Wales, 7. The gilgai soils. J. Soil Sci., 6:1-32.

Harris, S.A., 1959. The classification of gilgaied soils: some evidences from northern Iraq. J. Soil Sci., 10:27-33.

Harris, S.A., 1958. The gilgaied and bad structured soils of Central Iraq. J. Soil Sci., 9:169-185.

Hubble, G.D., R.F. Isbell and K.H. Northcote, 1983. Features of Australian soils. In Soils: an Australian Viewpoint. Division of Soils, CSIRO, Melbourne, pp. 17-47.

IBSRAM (International Board for Soil Research and Management Inc.), 1987. Management of Vertisols under semi-arid conditions. Proceedings of the First regional Seminar on Management of Vertisols under Semi-Arid Conditions, Nairobi, Kenya, 1-6 December 1986.

INWOSS, 1988. Transactions of the International Workshop on Swell-Shrink Soils, Nagpur, India.

ISSS, 1982. Vertisols and Rice Soils of the Tropics. Transactions, Symposium Papers II, 12th International Congress of Soil Science, New Delhi.

Kissel, D.A., J.T. Ritchie and E. Burnett, 1973. Chloride movement in undisturbed swelling clay soil. Soil Sci. Soc. Am. Proc. 37:21-4.

Knight, M.J. 1980. Structural analysis and mechanical origins of gilgai at Boorook, Victoria, Australia. Geoderma 23:245-283.

McGarity, J.W., E.H. Hoult and H.B. So, 1984. The properties and utilization of cracking clay soils. Reviews in Rural Science, No. 5, University of New England, Armidale, Australia.

Newman, A.L., 1986. Vertisols in Texas: some comments. USDA-SCS, Temple, Texas (Unpublished mimeographed draft.), pp. 206.

Paton, T.R., 1974. Origin and terminology for gilgai in Australia. Geoderma, 11:221-242.

Prescott, J.A., 1931. The soils of Australia in relation to climate, CSIRO Bull. 52.

Probert, M.E., I.F. Fergus, B.J. Bridge, D. McGarry, C.H. Thompson and J.S. Russell, 1987. The Properties and Management of Vertisols, CSIRO, CAB International, Wallingford, U.K.

Ritchie, J.T., D.E. Kissel and E. Burnett. 1972. Water movement in undisturbed swelling clay soil. Soil Sci. Soc. Am. Proc. 36:874-879.

Russell, J.S., A.W. Moore and J.E. Coaldrake, 1967. Relationships between subtropical semiarid forest of Acacia Harpophylla (Brigalow), microrelief, and chemical properties of associated gilgai soil. Aust. J. Bot., 15:481-98.

Soil Survey Staff, 1975. Soil Taxonomy. Soil Conservation Service, USDA, Handbook 436, 754 pp.

Soil Survey Administration, 1985. Taxonomy and management of Vertisols and Aridisols. Proceedings of the 5th International Soil Classification Workshop, Sudan, 2-11 November, 1982.

Spotts, J.W., 1974. The role of water in gilgai formation. Ph.D. Dissertation, Texas A&M University, College Station, Texas.

Stace, H.C.T., G.D. Hubble, R. Brewer, K.H. Northcote, J.R. Sleeman, M.J. Mulcahy, and E.G. Hallsworth, 1968. A Handbook of Australian Soils. Rellim Tech. Publ., Glenside, South Australia.

Stirk, G.B., 1954. Some aspects of soil shrinkage and the effect of cracking upon water entry into the soil. Aust. J. Agric. Res., 5:279-290.

Thompson, C.H. and G.G. Beckmann, 1982. Gilgai in Australian black earths and some of its effects on plants. Trop. Agric., 59:149-156.

Thompson, C.H. and G.G. Beckmann, 1959. Soils and land use in the Toowoomba Area, Darling Downs, Queensland. CSIRO, Aust. Div. Soils, Soils and Land Use Series No. 28.

Tucker, R.J., R.C. McDonald and M.D. Godwin, 1989. Soils of the Magoa River, Emerald irrigation area, Queensland. Land Resources Bull., Queensland Dept. of Primary Industries, Brisbane, Queensland, Australia. (In Preparation)

Wilding, L.P., 1985. Genesis of Vertisols. Proceedings of Fifth International Soil Classification Workshop, Sudan, 2-11 November 1982.

Wilding, L.P. and R. Puentes, (eds.), 1988. Vertisols: Their Distribution, Properties, Classification and Management., Texas A&M University Press.

Wilding, L.P. and J.A. Rehage. 1985. Pedogenesis of soils with aquic moisture regimes. Wetland Soils: Characterization, Classification, and Utilization. Proceedings of International Workshop. IRRI, Los Banos, Philippines, pp. 139-157.

Wilding, L.P. and D. Tessier, 1988. Genesis of Vertisols: shrink-swell phenomena. In L.P. Wilding and R. Puentes (eds.), Vertisols: Their Distribution, Properties, Classification and Management, Texas A&M University Press.

Yaalon, D.H. and D. Kalmar, 1978. Dynamics of cracking and swelling clay soils: displacement of skeleton grains, optimum depth of slickensides, and rate of intra-pedonic turbation. Earth Surface Processes, 3:31-42.

Yule, D.B. and J.T. Ritchie, 1980. Soil shrinkage relationships of Texas Vertisols, I: small cores. Soil Sci. Soc. Am. J. 44:1285-1291.

Geomorphology of Cold Deserts

Robert C. Palmquist[1]

Abstract

The cold desert is one in which at least one month has a mean temperature below 0°C so that frost action is possible. Most cold deserts occur in the mid latitudes as a result of continentality and/or orographic affects. The typical cold desert is of the mountain-and-basin type, wherein uplands of resistant rocks rise above a basin of less resistent rocks. These deserts are often tectonically unstable. The altitudinal differences give rise to through-flowing streams heading in the glaciers, snow fields, or forests of the more moist mountains.

The typical landforms of the cold desert are the pediment, stream terrace, alluvial fan, lake basin, and badlands. The repeated climatic and tectonic changes of the cold desert regions lead to sequences of relict landforms rising above small areas of active landscape development. The most common relict landforms are the pediment and terrace, each with a stable hillslope rising above it. Most of these relict surfaces increase in age with increasing height in multiples of 100,000 years, reflecting control by the climatic cycles of the Late Cenozoic.

Introduction

A desert is any naturally occuring area of sparse vegetation; as such there are lithological deserts on bare rock outcrops or gravelly substrates and climatic deserts in arid or very cold regions. This paper deals with deserts which are both arid and cold, although it avoids the extremely cold deserts of the polar regions. Cold deserts have at least one month with a mean air temperature of 0°C which is indicative of severe winters and frozen ground (Meigs, 1953). They are located in the middle latitudes and in the higher elevations of the low latitudes. In the mid latitudes, the aridity results from a combination of continentality, such as the deserts of Mongolia and the southern USSR, and from orographic effects, i.e. rain shadow deserts, as occur in North America. Bordering the deserts are extensive areas of semiarid climates which have similar landforms and soils (Figure 1).

Another important distinction between types of deserts is that of tectonism, i.e., the presence or absence of recent or ongoing geologic deformation. Most of the low-latitude deserts occur in areas of extreme stability of long duration. These *shield* or *platform* deserts are characteristic of Africa, Arabia, Australia, and India and are characterized by gently sloping erosional surfaces extending from granitic hills or table lands developed on either horizontal strata or weathering crusts of calcrete or silcrete.

In contrast, the mid-latitude deserts generally occur in regions of more active deformation with mountains rising above basins. These *mountain-and-basin* deserts are typified by the Basin and Ranges of Utah and Nevada and the intermontane basins of Wyoming and Colorado. Here, the elevational contrasts lead to climatic contrasts with more moist uplands contributing water from snow melt, or glaciers, or increased precipitation to streams which drain into or across the lowland deserts.

Another important distinction between the two types of deserts is that the basin-and-range desert has uplands composed of resistant rocks and basins composed of less resistant rocks. The combination of active or recent deformation and lithologic contrasts leads to a characteristic set of landforms.

This short paper attempts to make four points. First, the climate of all deserts has repeatedly changed with the climatic cycles of the Late Cenozoic. Second, the weathering processes of warm and cold deserts are similar except for the addition of frost action in cold deserts. Third, the variation in degrees of tectonic and climatic stability leads to the juxtaposition of relict and active landforms and to sequential landform development, which brings us to the fourth characteristic: most aridosols are likely to have developed under multiple climatic regimens. I will be unabashedly provincial in this paper and draw upon experiences and examples from the cold deserts and semiarid regions of the northern United States. This is an area with which I am familiar and through which you will be traveling on this trip.

[1]Northwest College, Powell, WY, 82435

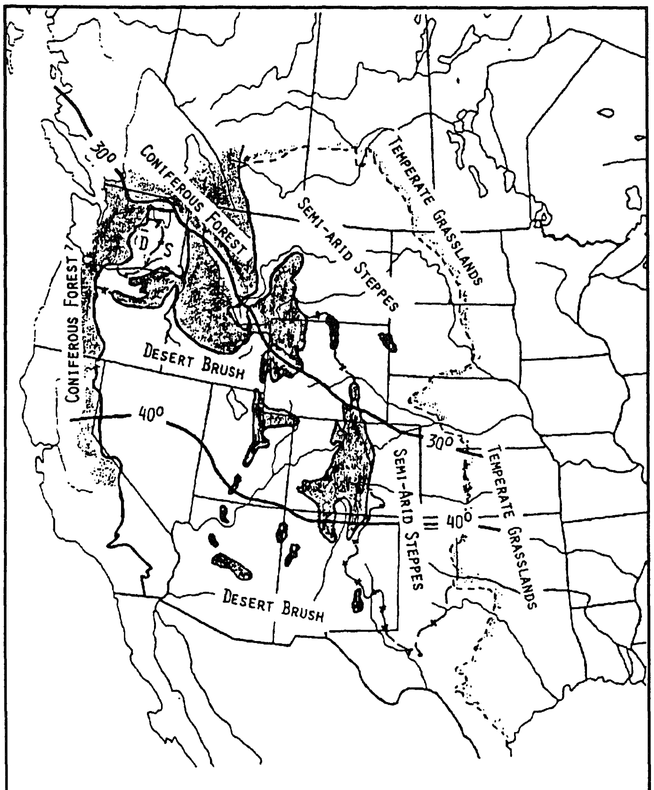


Figure 1. Location of deserts and semi-arid regions in North America as defined by vegetation. Cold deserts and semi-arid regions occur north of the 30°F isocline and warm deserts and semi-arid regions occur to the south of the 40°F isocline. A transition zone lies between. (Complied from several sources.)

Climatic Variations

The Late Cenozoic is characterized by repeated oscillations between colder and warmer and/or wetter and drier climates. The nature of these fluctuations is dependent upon the location of the desert. For instance, the warm deserts were subjected to an alternation of arid and savanna climates, which leads to the distinctive inselberg landscape developed by alternating chemical weathering and regolith stripping (Thomas, 1974). In contrast, the cold deserts have undergone an alternation of colder and warmer climates, which leads to the expansion and contraction of alpine glaciers in the adjacent mountains. The consequent variations in the load/discharge ratios of the mountain streams lead to the development of relict pediments, terraces, and alluvial fans.

Although the nature of the climatic change will vary with the particular setting of each desert, the timing of the changes is well described by the isotopic marine record. In this record can be read both the variations in mean sea surface temperatures and the volume of continental ice. The record indicates that the global climate has oscillated between colder and warmer (Figure 2) with approximately a 100,000-year frequency for the last 600,000 years and with a shorter, less regular frequency between 600,000 years and 2.5 million years, when the first evidence of extensive continental glaciations appears. During each cycle, the climate of an area varied greatly, as exemplified by the Fairbridge Cycle for mid latitudes (Figure 3a). The soils and landforms of the deserts reflect these climatic oscillations. The common perception that glacial climates were cooler and wetter than interglacial climates is an incorrect generalization. In the mid latitudes during maximal glaciation, the climate was colder but with about the same precipitation as at present (Figure 3b).

The cold deserts contain evidence of periglacial features such as ice-wedge casts, ground-wedges, and Mima-like mounds which are interpreted to have developed under arid to semi-

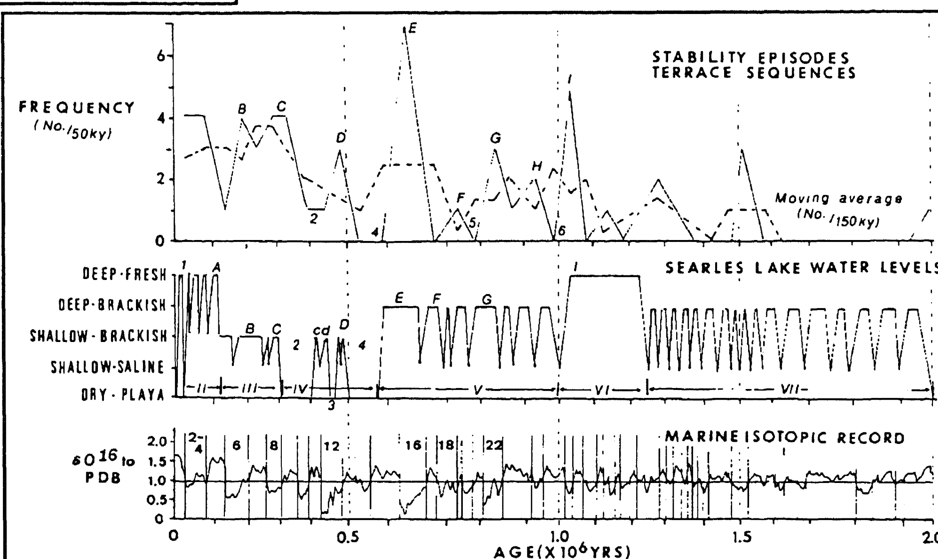


Figure 2. Geologic evidence of long term cyclic behavior. Bottom: The isotopic record of marine core V28-239; even numbered stages represent interglacials and odd numbered stage represent glacials (after Shackleton and Opdyke, 1976). Middle: Variations in salinity of Searles Lake, California based upon the sedimentology of a deep core (after Smith, 1984). Top: Frequency distribution of terrace ages along rivers in the Rocky Mountain West as estimated by the procedures in Palmquist (1983).

arid, windswept environments 10° to 13°C colder than today (about 5°C). (Mears, 1987). Ice wedge casts require a moist climate, such as that of northern Alaska, and ground wedges a drier climate, such as that of Anarctica.

Weathering Processes

The cold deserts are characterized by both physical and chemical weathering. The desert floors are littered with fragments of shattered rocks and sand from the granular disintegration of rocks. The shattering is the result of both insolation weathering and frost shattering, whereas the granular disintegration results from hydration, hydrolysis, and salt weathering perhaps accentuated by freezing. Chemical weathering processes such as oxidation, hydrolysis, solution, and leaching must exist or have existed in arid regions to explain such uniquely arid features as cavernously weathered boulders, duricrusts (silcretes and calcretes), and aridosols. Given the climatic variations of the Late Cenozoic, any weathering process could have operated at some time during the history of any desert landform or soil older than a few thousand years.

Landforms in Space and Time

The major landforms of the cold desert are the pediment, alluvial fan, stream terrace, and pluvial lake basin. These piedmont features border hillslopes developed on either the resistent rocks of the mountain flanks or the badlands of the basinal uplands and reflect differences in lithology and/or tectonic stability and climatic variations. Their present activity influences the distribution of aridosols, inceptisols, and entisols.

Pediments

The major erosional feature of the piedmont is the pediment (Figure 4) which is best defined as a bedrock erosional surface of low relief extending basinward from a backing hillslope from which it is separated by an abrupt change of gradient (freely modified from Marbutt, 1977). Most pediments in cold deserts are mantled with debris from the adjacent uplands and are termed *mantled pediments*. An exception is the relatively small badlands pediment, which is free of a mantle and is properly termed a *pediment*. Mantled pediments are generally concave up in longitudinal profile and have a steeper gradient than stone-free pediments. In transverse section, pediments undulate with low interfluves separating drainage lines. The sediment on the interfluves may be less well sorted than that along the drainages.

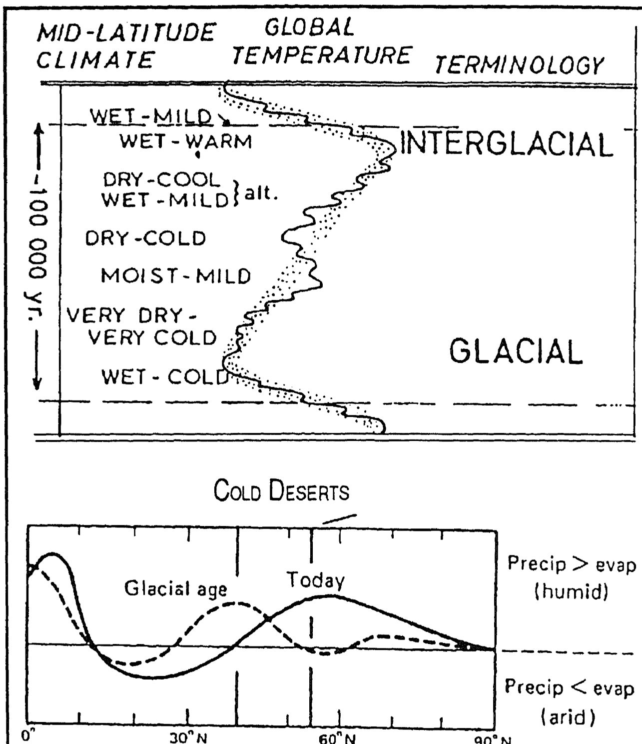


Figure 3. Top: Fairbridge Cycle of climatic change within a single interglacial-glacial cycle as defined by one odd and even numbered isotopic stage (Fig. 2). Cycle based upon a mid-latitude, maritime climate and illustrates the complex variation in moisture and temperature (simplified from Fairbridge, 1972). Bottom: Effective variation of precipitation in the northern hemisphere between an interglacial (today) and glacial illustrating the latidutinal changes of the cold deserts during a glacial cycle. Soild line represents present conditions; broken line indicates glacial conditions. The vertical lines mark the latidutinal range of present day cold deserts. Diagram does not consider the affects of rain shadow in the latidutinal distribution of desertic conditions. (Modified from Flohn, 1953.)

All hypotheses to explain the genesis of pediments demand that they form in an environment with a stable baselevel. This limitation restricts the formation of large pediments to areas of tectonic and climatic stability. Given the climatic fluctuations of the Late Cenozoic, it is not surprising that most pediments in cold deserts are relict and are presently being dissected. In my experience, the youngest relict pediments have ages which place them in one of the last three interglacials, that is, in isotopic stage 5, 7, or 9 (Figure 2), with the youngest mountain front pediments generally falling into

stage 7 or 9 and the youngest badlands or basin pediments falling into stages 5 or 7.

Badlands pediments which develop rapidly through slope retreat in the soft clays may be either active or relict. In any extensive badlands area, most of the larger pediments are relict and stand above the modern drainage. Only the smaller pediments that border and are graded to the modern drainage are actively forming. The active pediments have developed since the end of the Pleistocene about 11,000 years ago, and most are related to the end of the Altithermal about 3,000 years ago. The inactive pediments are older than stage 5 and are bordered by inactive back slopes.

Alluvial Fan

The *alluvial fan* is found in areas of active stream incision which may result from active tectonism, recent climatic change, or master stream incision. Alluvial fans are constructed from deposition either from streams or debris flows or a combination of the two. Generally, the more arid the region, the more dominant the deposition from debris flows. Debris flows result from the supersaturation of the stream by debris eroded from the valley floors of the uplands. On the more permeable fan, the interstitial water is lost and the lobate flow comes to rest. Generally the flow deposit is coarse-grained and poorly sorted and is confined to the channel or its immediate environs. Alluvial deposits are more heterogeneous and consist of finer-grained overbank deposits cut by linear sandy or gravelly channel deposits.

Most alluvial fans are concurrently being built by deposition while at the same time they are being destroyed by erosion (Figure 5). The active portion of the fan is characterized by exposed silts, sands, and gravel deposited from debris flows or streams. This portion of the fan is generally at a slightly higher elevation than the stable or actively degrading portions. The stable portions are mantled by a one-stone thick layer of gravel, *desert pavement*, which is lag from erosion by wind or overland flow of runoff. The longer that the surface has been stable, the better developed the pavement and its dark coating of manganese dioxide, *desert varnish*. The portions of the fan being degraded contain numerous gullies which head upon the fan surface.

This complex activity occurs within one depositional cycle. Given the repeated climatic changes of the Late Cenozoic and the repeated alternations of depositional and erosional cycles engendered by it, any fan surface is likely to have a wide variety of Aridisols and Entisols.

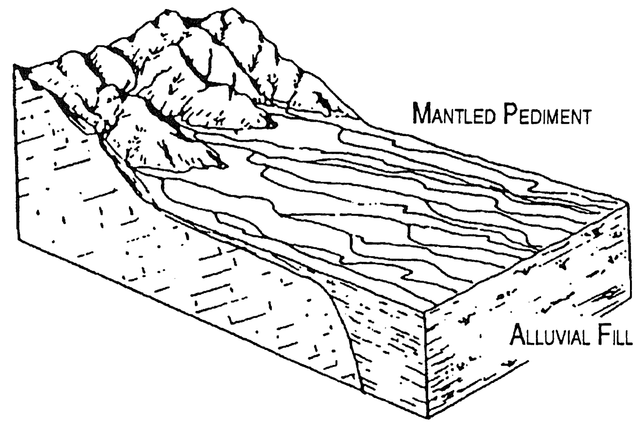


Figure 4. Block diagram illustrating the features of a mantled pediment. The thick alluvium at its base either may be the product of deposition within a closed basin or be shallower in depth and the product of deposition of stable stream in an open basin. (Modified from Packard, 1974, as presented in Selby 1982.)

Stream Terraces

Stream terraces are flood plains abandoned by the rapid incision of the channel. As such, they stand above the modern flood plain as a bench and testify to the increase in erosional power of the channel at some time in the past. Terraces are common in the mountain-and-basin deserts because an increase in erosional power will result from (1) increased gradient - perhaps as the result of tectonic tilting or incision of the master stream, (2) increased discharge - probably the result of a climatic change, or (3) decreased load - most likely the product of a climatic change. The Late Cenozoic variations in glacial activity caused the mountain-bred streams to have repeated variations in their load/discharge ratios, which lead to alternating episodes of stability and incision.

Most available ages for geomorphic features suggest that glacial-age stability or deposition alternated with interglacial-age incision, as exemplified by the terrace sequences along the rivers in the Rocky Mountain West (Figure 2). The result of such alternations is a chronosequence grading from Typic Calciorthids on the lowest terrace to Ustollic Haplargids on the higher terraces but with gradations in soil development not reflected in the terminology changes (Reheis, 1987a, 1987b). Any hillslope,

pediment, or alluvial fan graded to a terrace is likewise relict and with an equivalently developed soil.

Desert Lake Basins

Within many of the closed basins of the mountain-and-basin deserts of Utah and Nevada, fresh water lakes developed during the cooler glacial stages. Some of these lakes, such as glacial Lake Bonneville at Salt Lake City, were very extensive and deep and developed around their shorelines deltas, beaches, wave-cut cliffs and benches. These lakes varied from fresh water to highly saline, depending upon the relationship between the elevation of the water surface and that of the outlet. This relationship is best determined from the composition of the sediments (Figure 2). Within each cycle, as the water level fell, a series of shorelines were exposed, on which have developed a complex chronosequence.

Desert Hillslopes

Bordering the piedmont features are hillslopes which range from the debris mantled or bare rock slopes of the mountain front to the badlands slopes of the basins. The slopes vary greatly in their vegetation cover and stability. Many of the mountain front slopes are either exposed limestones, sandstones, or granites, with scattered shrubs growing from joints, or debris-mantled with scattered shrubs and grasses. Most of these slopes are stable and lead down to relict pediments at their bases.

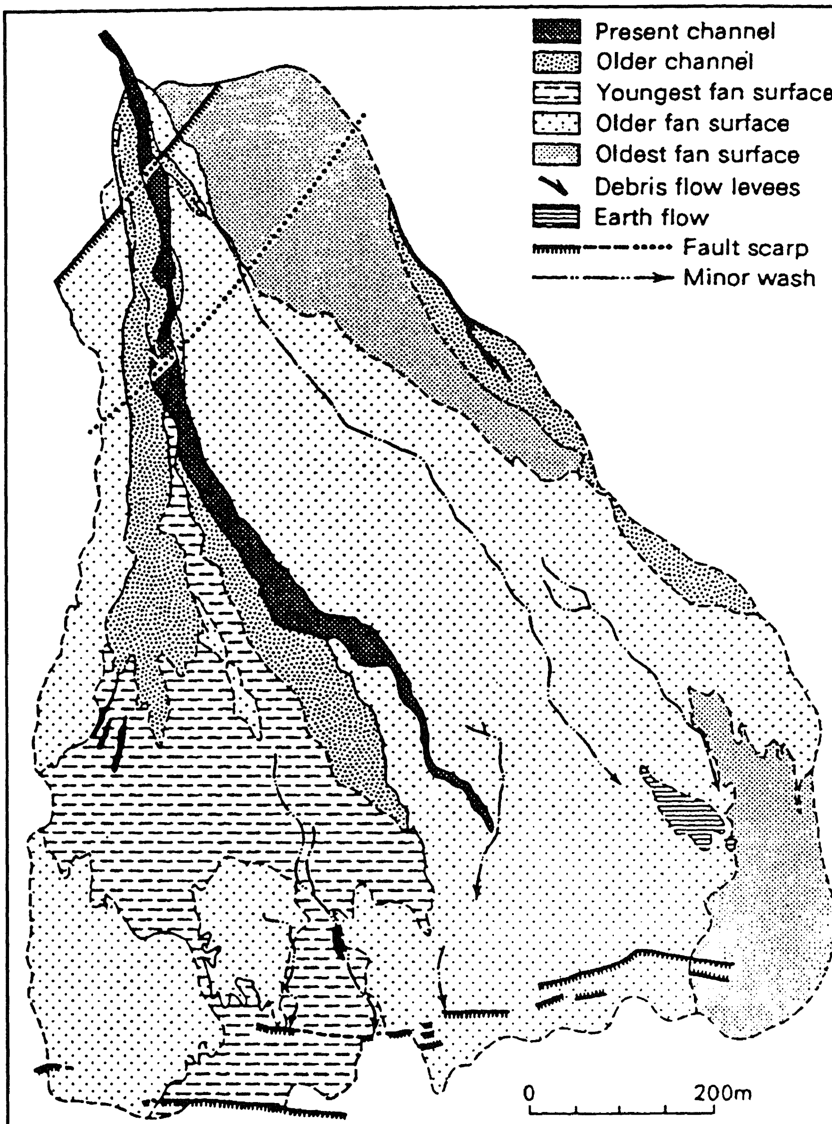


Figure 5. Variation in surface ages and processes on the Shadow Rock Fan, California illustrating the localized nature of deposition. The older surfaces will have increasingly well developed desert pavement, desert varnish, and aridosols (after Hooke, 1967).

Within the basins, hillslopes flank relict pediments, alluvial fans, and terraces or form the intricately dissected mosaic of the badlands. Those slopes flanking relict piedmont features are usually gentler, more subdued, and more vegetated than the slopes of the badlands and have vegetated drainages at their base. It is difficult to escape the conclusion that they are likewise relict and that the only active slopes are those of the badlands.

Badlands occur in three situations. The most common is at the head of actively incising tributaries leading to a through flowing master stream. Generally these tributaries have breached terraces and alluvial fans to dissect portions of the otherwise stable slopes rising above these features. The second situation is at the head waters of major basinal streams where steep slopes exist because of cap rocks, and the third is where a stream of any size is undercutting the valley wall at the outside of a meander.

The distribution of badlands indicates that the dissection of a hillslope is controlled by the activity of its base. A hillslope rising above a terrace, relict pediment, or relict alluvial fan is stable, whereas a slope rising above an incising base is actively being degraded. In my experi-

ence, few degrading hillslopes have stable bases upon which colluvial slopes or small alluvial fans are being constructed. The implication of this observation is that the present arid environment is incapable of slope dissection unless aided by slope oversteepening through either basal erosion or height exaggeration resulting from a cap rock.

Spatial and Temporal Variability

As has been pointed out in the previous paragraphs, the climate and/or tectonism of the cold deserts has undergone repeated changes throughout the Late Cenozoic. The result is that these deserts are a complex mosaic of relict and active surfaces. The relict surfaces, such as pediment and stable backslope, rise as a series of benches above the modern drainage with ages that generally increase with height in multiples of 100,000 years. Between the sets of stable surfaces are actively degrading badlands, perhaps containing landslides, along the margins of and in the headwaters of incising streams. All of the incising streams are tributary to a through-flowing, mountain-bred, master stream.

The result of this spatial juxtaposition of varible-age surfaces is a mosaic of soils which vary in a complex manner. The variation in soils results from differences in parent materials, topographic position, microclimate and vegetation, and age. In many respects within the cold desert, age is the most important of all of the soil forming factors. Age influences not only the degree of development of the soils but also the number of climatic cycles through which the soil has developed. Alternation of climatic cycles leads to such features as carbonate-engulfed, argillic horizons and frost shattered, petrocalcic horizons, corrosion surfaces within carbonate nodules, and the stripped and redeposited epipedon of the Paleargid. All of these characteristics require a fluctuation between either colder and warmer climates and/or wetter and drier climates.

One might properly ask the questions — Which characteristics of the cold desert aridosol are the product of the environment in which they are presently found? What landforms of the cold deserts were formed under the present environment? Did the arid characteristics of the cold desert aridosol form under interglacial-age aridity or glacial-age aridity?

The cold desert and its soils are a complex ensemble with distinctive characteristics formed by its unique history of climatic and tectonic variations. Its similarities to the warm deserts are obvious, but their different geologic, climatic, and hence pedogenic histories make them different entities.

References Cited

Denny, C.S. 1967. Fans and Pediments. Am. J. Sci. 265:81-105.

Fairbridge, R.W. 1982. Climatology of a glacial cycle. Quat. Res. 2:283-302.

Flohn, H. 1953. Studien uber die atmosphariische Zirkulation in der letzten Eiszeit. Erdkunde 7:266-275.

Hooke, R. leB. 1967. Processes on arid-region alluvial fans. J. Geol. 5:438-60.

Mabbutt, J.A. 977. Desert Landforms. MIT Press, Cambridge.

Mears, B., Jr. 1987. Late Pleistocene periglacial wedge sites in Wyoming: an illustrated compendium. Menoir No. 3. The Geological Survey of Wyoming, Laramie.

Meigs, P. 1953. World distribution of arid and semi-arid homoclimates. Rev. of Res. on Arid Zone Hydrology. Arid Zone Programme 1. UNESCO, Paris. 203-210.

Palmquist, R.C. 1983. Terrace chronologies in the Bighorn Basin, Wyoming. 34th Ann. Field Conf., Wy. Geol. Assoc. Guidebook. 217-231.

Reheis, M.C. 1987a. Soils in granitic alluvium in humid and semiarid climates along Rock Creek, Carbon County, Montana. US Geol. Sur. Bull 1590.

Reheis, M.C. 1987b. Climatic implications of alternating clay and carbonate formation in semiarid soils of south-central Montana. Quat. Res. 27:270-282.

Shackleton, N. J. and N. D. Opdyke, 1973. Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale. Quat. Res. 3:39-55.

Smith, G.I. 1984. Paleohydrologic regimes in the south-western Great Basin 0-3.2 my ago, compared with other long records of "global" climate. Quat. Res. 22:1-17.

Selby, M.J. 1982. Hillslope materials and processes. Oxford University Press.

Thomas, M.F. 1974. Tropical Geomorphology. John Wiley and Sons, New York.